GEOLOGICAL SURVEY OF NEWFOUNDLAND
DEPARTMENT OF MINES
GOVERNMENT OF NEWFOUNDLAND AND LABRADOR

CURRENT RESEARCH

REPORT 89 – 1

Compiled and Edited by
C.P.G. Pereira, D.G. Walsh and R.F. Blackwood

St. John's, Newfoundland
March, 1989

Geological Survey of Newfoundland
Department of Mines
Government of Newfoundland and Labrador

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PREFACE

This publication continues the series of annual reports initiated by the Mineral Development Division of the Newfoundland Department of Mines in 1974. The forty reports contained herein present the preliminary results of the 1988 research program. Final results will be published later, as individual reports on each project.

The Mineral Development Division, which has in fact functioned as a geological survey since 1972, was renamed the Geological Survey of Newfoundland and raised to Branch status in a major reorganization of the Newfoundland Department of Mines, late in 1988. This report is the first Current Research published under the new name. It marks the reinstatement of the Geological Survey of Newfoundland, originally established in 1864 under Alexander Murray, and a return to a publication series bearing that name after a lapse of thirty years.

Many of the projects described herein were funded under the Canada–Newfoundland Mineral Development Agreement (1984-1989). The publication also includes abstracts of reports on work carried out under the Agreement by the Geological Survey of Canada. These reports will be published in the GSC Current Research (GSC Rept. 89-1); we thank GSC for permission to publish the abstracts here.

B.A. Greene
Assistant Deputy Minister
Geological Survey of Newfoundland

Any reader who would like to write a rebuttal to, or discussion of, any report contained in this volume is invited to submit it to the editor by November 1, 1989, to be considered for inclusion in Report 90−1.

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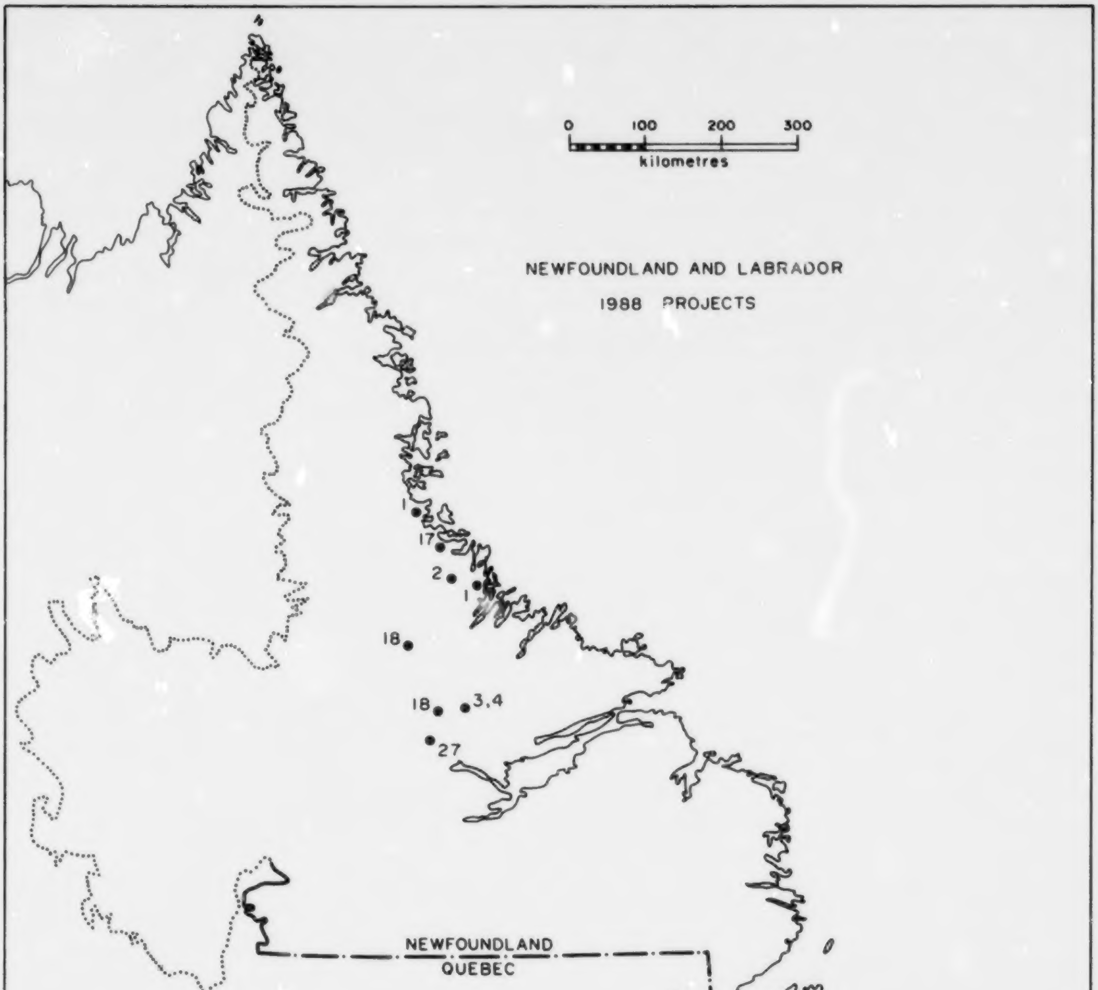
NEWFOUNDLAND AND LABRADOR
1988 PROJECTS

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NEWFOUNDLAND
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THE PETROGENESIS AND EMPLACEMENT OF PROTEROZOIC DYKE SWARMS, PART 1: INITIAL FINDINGS AND PATHS OF FURTHER RESEARCH

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ABSTRACT

Proterozoic dyke swarms intruding Archean shield terranes represent major events of igneous activity in the Earth's history. Recent debate has focussed on the role of source heterogeneity versus that of crustal contamination, in order to explain geochemical discrepancies between major- or trace-element compositions as opposed to isotope abundances. The emplacement of Proterozoic dyke swarms constitutes an episode of crustal addition, in previously permobile environments, that may record, or herald, fundamental changes in global tectonic processes. Study of these dyke swarms may reveal if crust-manufacturing processes were similar to, or different from, those of the Phanerozoic and how, if at all, they changed through time.

INTRODUCTION

The unique scale and frequency of the dyke swarms that were intruded into newly stabilized Archean cratons from Late Archean to Late Proterozoic time, make these dykes excellent sources for studying the geochemical evolution of dyke swarms and their mantle source, during this same period. The Kikkertavak dyke swarm and the Harp Lake dykes, emplaced into the Archean Hopedale Block of Labrador, were chosen to test the application of existing models pertaining to swarm magma composition and mantle source changes during the Late Archean to Middle Proterozoic. In addition, the direction of magmatic flow and swarm emplacement processes are examined. Geochronological studies have shown the Kikkertavak dykes to be about 2200 Ma (Rb−Sr whole rock date, B.J. Fryer, *in* Grant *et al.*, 1983), and the Harp dykes to be about 1200 Ma (Rb−Sr whole rock date, F.R. Voner, *in* Ermanovics *et al.*, 1982; Grant *et al.*, 1983).

PREVIOUS WORK

Little detailed research has been done on the Hopedale Block. The present evolutionary history of the area is based on regional 1:100,000-scale mapping by the Geological Survey of Canada (GSC) (Ermanovics, 1979, 1980, 1981a,b 1984; Ermanovics and Raudsepp, 1979; Ermanovics and Korstgård, 1981; Ermanovics *et al.*, 1982). The GSC focus has been to establish the petrological, geochronological and structural aspects of the Archean gneisses (e.g., Korstgård and Ermanovics, 1984, 1985). Earlier work in the area has also concentrated on structural and metamorphic histories (e.g., Taylor, 1971) and correlation with similar terrains in west Greenland (e.g., Sutton *et al.*, 1972).

Apart from brief descriptions in the GSC reports (see above), very little data are available on the Kikkertavak dykes. The Harp dykes, a series of younger intrusions within and adjacent to the Harp Lake anorthosite−gabbro plutonic complex approximately 100 km west of the present study area, have been investigated by Meyers and Emslie (1977).

GEOLOGICAL SETTING OF THE DYKE SWARMS

The Hopedale Block lies in the southern part of the Nain Province, adjacent to the Churchill Province in the west and the Makkovik Province in the southeast (Figures 1 and 2). It consists of polydeformed, mainly granodioritic to tonalitic gneisses having a predominantly northeast−southwest (Fiordian) trend. The Fiordian-trending fabrics are interpreted by Ermanovics *et al.* (1982) to be Late Archean and postdate the predominantly pretectonic Kanairiktok granite at about 2830 Ma (U−Pb zircon concordia, Loveridge *et al.*, 1987). The Fiordian trend overprints an earlier northwest−southeast (Hopedalian) trend that is preserved around and to the northwest of the village of Hopedale (Figure 2) and which gives U−Pb zircon and Rb−Sr whole rock dates in the range between 3250 and 3000 Ma (Grant *et al.*, 1983; Loveridge *et al.*, 1987). Within both the Fiordian and Hopedalian domains, subvertical planar and linear fabrics show strong parallelism, suggesting block-wide, shear-zone deformation in both cases. The Fiordian-trending gneisses, which dominate the study area, are deformed agmatitic gneisses having some amphibolite bands. The gneisses are at amphibolite grade, possibly retrogressed from a granulite-facies assemblage. Any earlier structural and metamorphic histories were largely erased by the Hopedalian event (Ermanovics *et al.*, 1982).

The post-Fiord-trend dykes are very conspicuous as rectiplanar and irregular intrusive masses throughout the

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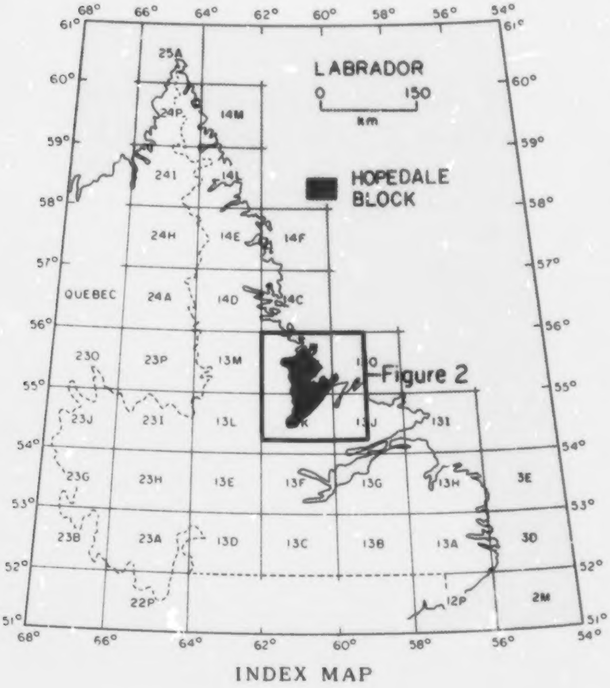
[1] This research is part of the International Geological Correlation Programme (IGCP) Project Number 257 on Precambrian Dyke Swarms.


LABRADOR
0 150
km
HOPEDALE BLOCK
QUEBEC
Figure 2
25A
24P
14M
24I
14L
24H
14E
14F
24A
14D
14C
23O
23P
13M
13O
23J
23I
13L
13K
13J
13I
23G
23H
13E
13F
13G
13H
3E
23B
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53°
52°
51°
INDEX MAP

Figure 1. *Location of the Hopedale Block, Labrador.*

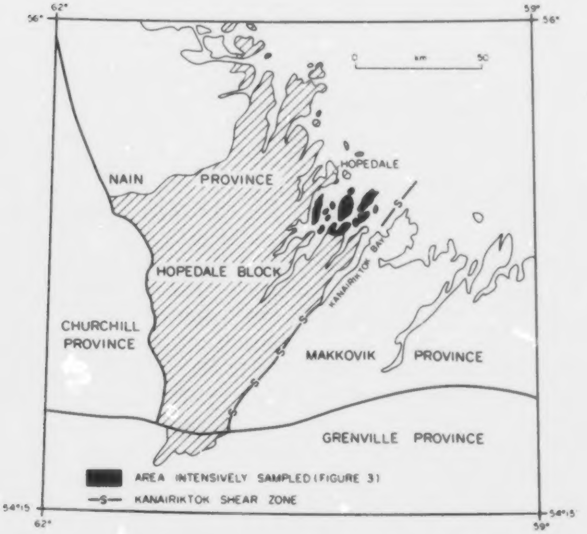
62°
59°
56°
54°15
0 km 50
HOPEDALE
NAIN PROVINCE
HOPEDALE BLOCK
KANAIRIKTOK BAY
CHURCHILL PROVINCE
MAKKOVIK PROVINCE
GRENVILLE PROVINCE
AREA INTENSIVELY SAMPLED (FIGURE 3)
—S— KANAIRIKTOK SHEAR ZONE

Figure 2. *The study area within the Hopedale Block.*

whole Hopedale Block. They are superbly exposed along the ice-scoured coastline of the mainland and island archipelago, north of Kanairiktok Bay (Figure 3). Consequently, this area was chosen for detailed investigation, and fieldwork was undertaken mainly by boat. Wherever possible, centres, margins and wallrock contacts of dykes were sampled for geochemical analysis. Studies of dyke emplacement concentrated on very detailed mapping of primary dyke structures and contact kinematic indicators.

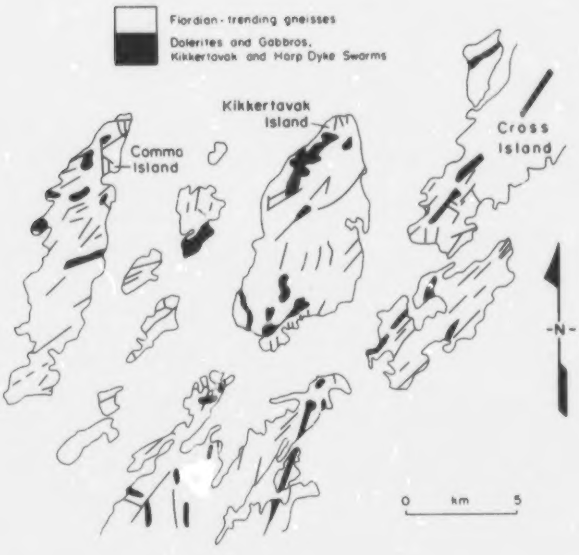
Fiordian-trending gneisses
Dolerites and Gabbros, Kikkertavak and Harp Dyke Swarms
Kikkertavak Island
Commo Island
Cross Island
-N-
0 km 5

Figure 3. *Sketch map showing dyke trends in the intensively sampled area.*

GEOCHEMISTRY AND PETROLOGY

The overall character of the two swarms appears very similar in hand specimen: the older Kikkertavak swarm consists of clinopyroxene–plagioclase diabase and olivine–pyroxene–plagioclase gabbro and the younger Harp dykes consist of olivine diabase and olivine gabbro (Ermanovics *et al.*, 1982).

In the field, the most diagnostic feature of the diabases in each swarm is colour. The low-grade, greenschist-facies alteration (and rare, localized shearing of the margins) of the Kikkertavak dykes, caused by northeast–southwest shearing and deformation in the mid-Proterozoic, e.g., the Kanairiktok shear zone (Ermanovics *et al.*, 1982), has led to a light green colouration, which is especially widely developed in the southern part of the Hopedale Block. The Harp dykes, however, often have a deep red-spotted, brown surface, due to their high iron content and the weathering of olivine crystals respectively, but appear unaltered. The gabbros, on the other hand, are seemingly indistinguishable and cannot be confidently assigned to a particular swarm on the basis of field criteria.

The locally very strong plagioclase phyricity of the Kikkertavak swarm, totally absent in the Harp dykes, points to considerable plagioclase fractionation within these dykes. Phenocrysts measure up to 10 cm in diameter and are usually euhedral or subhedral in shape, although a few multi-nucleic aggregates and porphyritic, cognate xenoliths are also present.

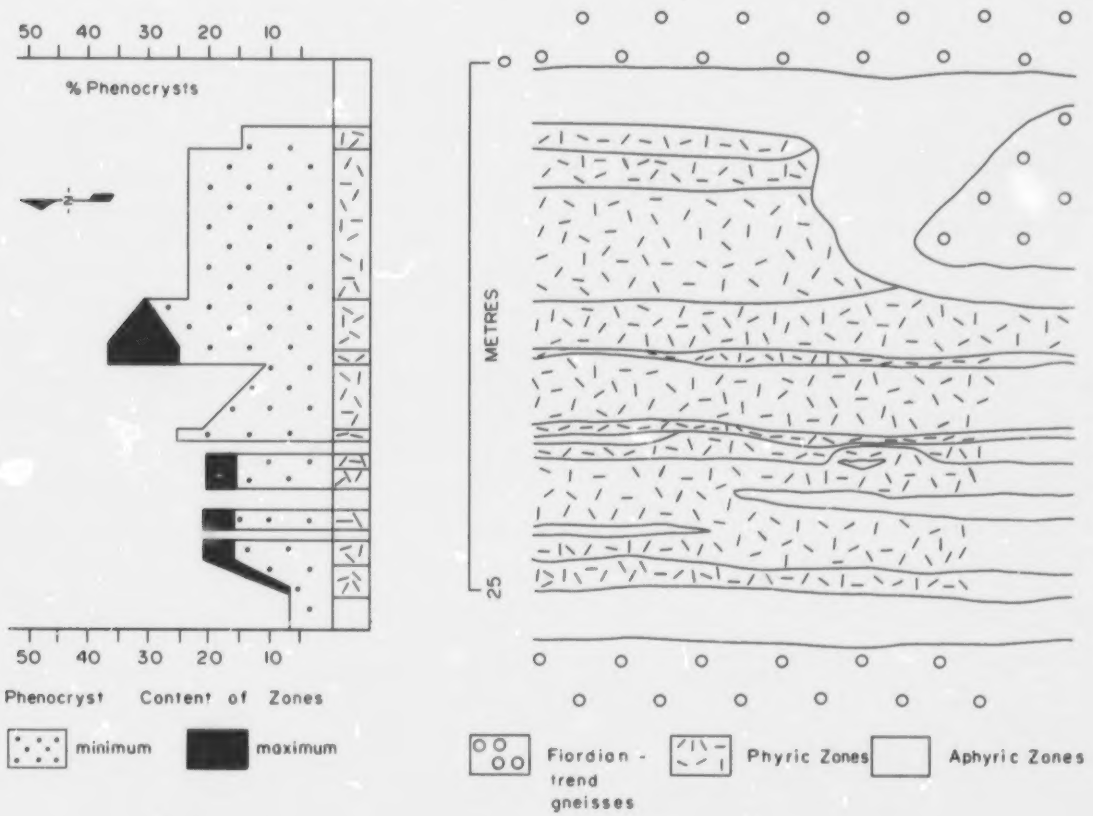

Figure 4. *Diagram showing marked zoning of phenocrysts in a Kikkertavak dyke, Cross Island.*

Reaction rims are absent, suggesting equilibrium with the surrounding melt. Division into phyric and aphyric zones within a dyke is commonly marked. Commonly, the phenocrysts are highly concentrated in the central portion of the dyke, suggesting flow differentiation (Komar, 1972). However, one particular dyke shows a very complex, layered, zoning pattern (Figure 4), suggesting that differential accretion caused the resulting zoning pattern (Platten and Watterson, 1987). The differential accretion theory argues that changing hydraulic conditions in the magma were responsible for the accretion or non-accretion of phenocrysts onto the advancing solid–liquid interface in the dyke: phenocryst concentration in the dykes need not change during intrusion and the magma is porphyritic throughout emplacement.

Dyke margins are chilled and no visible phenocrysts are seen accumulated to them. Their absence indicates that heat transfer was by conduction rather than by convection (Tarney and Weaver, 1987) and concurs with the differential accretion of phenocryst phases mentioned above. Convectional heat transfer would be expected to deposit coarse phenocrysts on dyke margins or to result in the thermal erosion of the country rock. Also, in the absence of wallrock cumulates, a convecting basaltic magma would partially melt gneisses of granodioritic or tonalitic composition, to produce selective contamination of the magma (Mohr, 1987). In contrast, conduction would be expected to insulate the country rock (granodioritic–tonalitic gneiss) as rapid cooling took place at the dyke–wallrock interface. Also, no evidence of corrosion is seen in the field: primary margin morphology (bayonet and step features) is preserved, and country rock or xenolith contacts with dykes are always sharp. Weaver and Tarney (1981) suggest that a lack of crustal contamination is a general feature of many Proterozoic dyke swarms.

Despite the lack of contamination, the heavy zoning of phenocrysts in the Kikkertavak swarm means that a primary magma composition is unlikely to be preserved in either the centre or the margins (Weaver and Tarney, 1983). The Harp dykes, however, are not visibly differentiated and, hence, samples may preserve the original magma geochemistry.

Therefore, the best approximation of a primary magma is not necessarily the most primitive of analyzed samples. Although work on the Scourie dykes (Weaver and Tarney, 1983) suggests that shallow level processes, (e.g., fractionation or mixing) may not affect many Proterozoic swarms, differentiation across the dykes complicates modelling. The trend of progressive magmatic evolution from the dyke margin to the centre as the fissure dilates with time is, therefore, complicated in the Kikkertavak dykes. Analyzing only the groundmass in these cases may obtain a clearer trend.

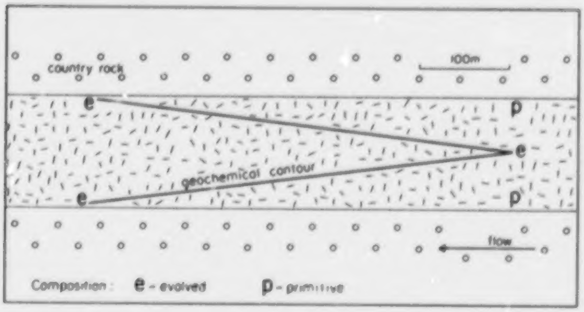
country rock
100m
e
p
e
geochemical contour
e
p
flow
Composition : e - evolved p - primitive

Figure 5. *Analysis of magma flow by geochemical contouring.*

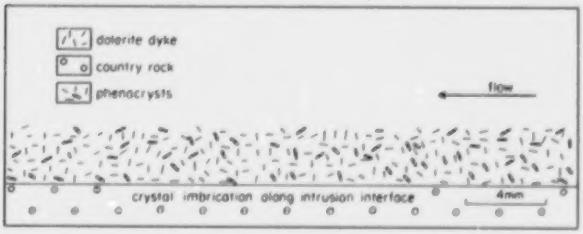
dolerite dyke
country rock
phenocrysts
flow
crystal imbrication along intrusion interface
4mm

Figure 6. *Analysis of magma flow by crystal imbrication.*

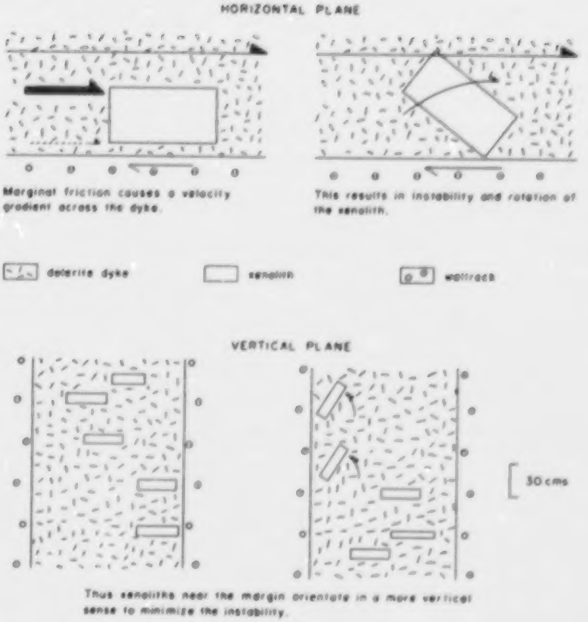
HORIZONTAL PLANE
Marginal friction causes a velocity gradient across the dyke.
This results in instability and rotation of the xenolith.
dolerite dyke
xenolith
wallrock
VERTICAL PLANE
30 cms
Thus xenoliths near the margin orientate in a more vertical sense to minimize the instability.

Figure 7. *Three-dimensional representation of xenolith orientations in a Harp(?) dyke, Cross Island.*

MAGMA FLOW AND DYKE EMPLACEMENT

Magmatic flow within the dykes, during intrusion, is one of the main objects of study of the IGCP research.

The Hopedale crustal block is not considered to have been tilted significantly after emplacement of the swarms: Fiordian metamorphic grade is fairly constant throughout. (However, the presence of low-angle mafic sheets in the north of the region, suggests a shallower crustal level may be exposed there.) Therefore, a component of lateral flow within the dykes should be apparent as a progressive evolution of magma composition along the length of a dyke (Figure 5). Flow direction can also be indicated by crystal or xenolith alignment within a dyke, especially if imbricated along wallrock contacts (Figure 6). This latter criterion would also give a monopolar, rather than a bipolar, flow direction.

Generally, dykes are free of xenoliths, but at one remarkable locality (Figure 7), 60 percent of a dyke consists of tabular xenoliths. This dyke quickly dies out vertically, suggesting that the xenoliths are the product of roof stoping. However, many xenoliths are not composed of locally indigenous granodioritic gneiss, but amphibolite or metaquartzite, suggesting derivation from elsewhere. The xenoliths are orientated with the largest plane striking parallel to the dyke wall and suggest horizontal flow. The xenoliths are usually horizontal in the centre of the dyke but subvertical at the northern margin (Figure 7). This change is still consistent with lateral flow since the dyke wall would exert a frictional shear force, and a horizontal xenolith adjacent to the dyke wall would have a stronger force applied to its outer edge by the magma flow. Thus, a moment of force would be created, rotating the xenolith around the 'a' axis (Figure 8). This instability could be minimized by orienting the xenolith parallel to the dyke wall, therefore, eradicating the differential force (Figure 8). The southern margin does not show this phenomenon due to the presence of an extremely large xenolith block (greater than one metre in length) that caused local turbulence in the flow.

The general lack of xenoliths in most dykes suggests dilation to be the mechanism of emplacement. The direction in which this dilation occurs depends on three factors; 1) pre-intrusive weaknesses in the host rock, 2) the regional stress field, and 3) the local stress field induced by magma pressure. If host rock weaknesses are exploited, then a dyke is probably not intruding along a principal stress direction resulting in an oblique dilation (Delaney *et al.*, 1986). The extrapolated closure of a Harp dyke in the horizontal plane (Figure 9), suggests a dilation direction perpendicular to the dyke wall, and if further studies on both swarms indicate the same, then pre-existing lineaments will not have influenced dyke emplacement. Therefore, emplacement involves interaction between regional and magmatic stresses. Nearer the centre of magma pressure, opening direction may be largely controlled by it, whereas at distal regions of the swarm, magma pressure wanes and the regional stress field will increase in influence (Odé, 1957).

This interaction is difficult to analyze in three dimensions, as both horizontal and vertical planes of the dyke must be examined. For example, Figure 9 shows a carefully

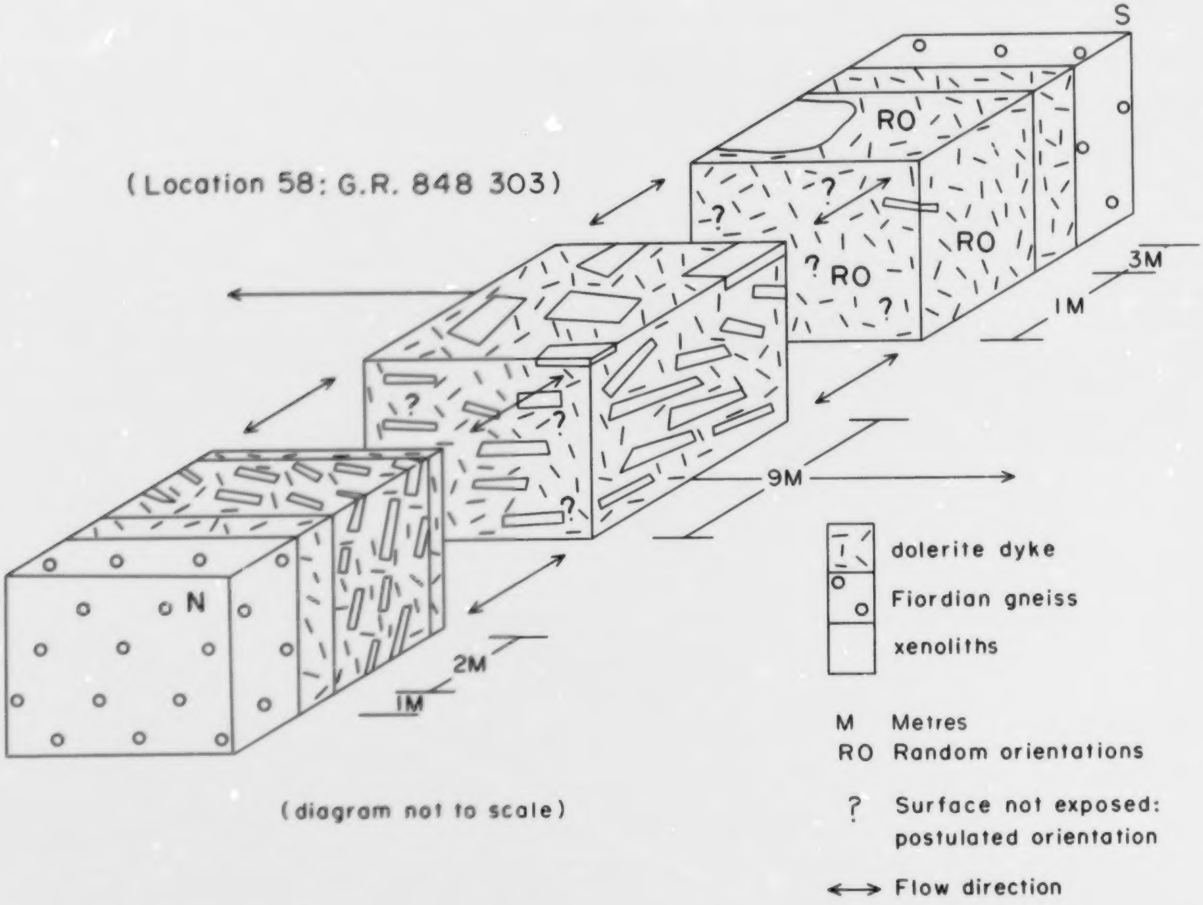
(Location 58: G.R. 848 303)
S
N
RO
?
3M
1M
9M
2M
1M
(diagram not to scale)
dolerite dyke
Fiordian gneiss
xenoliths
M Metres
RO Random orientations
? Surface not exposed: postulated orientation
Flow direction

Figure 8. *Stress differentials at dyke margins and potential resultant rotations.*

measured Harp dyke mapped in the horizontal plane. The extrapolated horizontal closure leaves 15 percent of the dyke surface area unaccounted for. Corrosion of the gneissic wallrocks to this degree is highly unlikely as already discussed, therefore, some degree of dilation in the vertical plane must have occurred.

THE NATURE OF THE PROTEROZOIC MANTLE

The large scale of Proterozoic dyke swarms, points to the mantle as the original magma source. Work on the Scourie dykes (Weaver and Tarney, 1983), suggests that geochemical heterogeneities in the swarm cannot be modelled by shallow level fractionation or contamination, but evolve from heterogeneities in a previously metasomatized mantle. Although the Kikkertavak and Harp dyke swarms appear similar and homogeneous, the degassing and depletion of the mantle during the 1 Ga period between their respective intrusions may be reflected in their subsequent geochemistry.

FUTURE PLANS

The 338 samples taken during the last field season (1988) for sectioning, geochemical analysis and dating, should be sufficient for petrological studies. However, because the sampling was intensive but regionally restricted (Figure 2), further sampling may be necessary in outlying parts of the field area in order to analyze regional trends in swarm geochemistry and petrology.

In the summer of 1989, further work will focus on dyke emplacement processes, especially on their relationship to pre-intrusive foliations and joint patterns within the host gneisses. Fissure dilation directions may reveal information concerning the paleostress system operating during dyke emplacement.

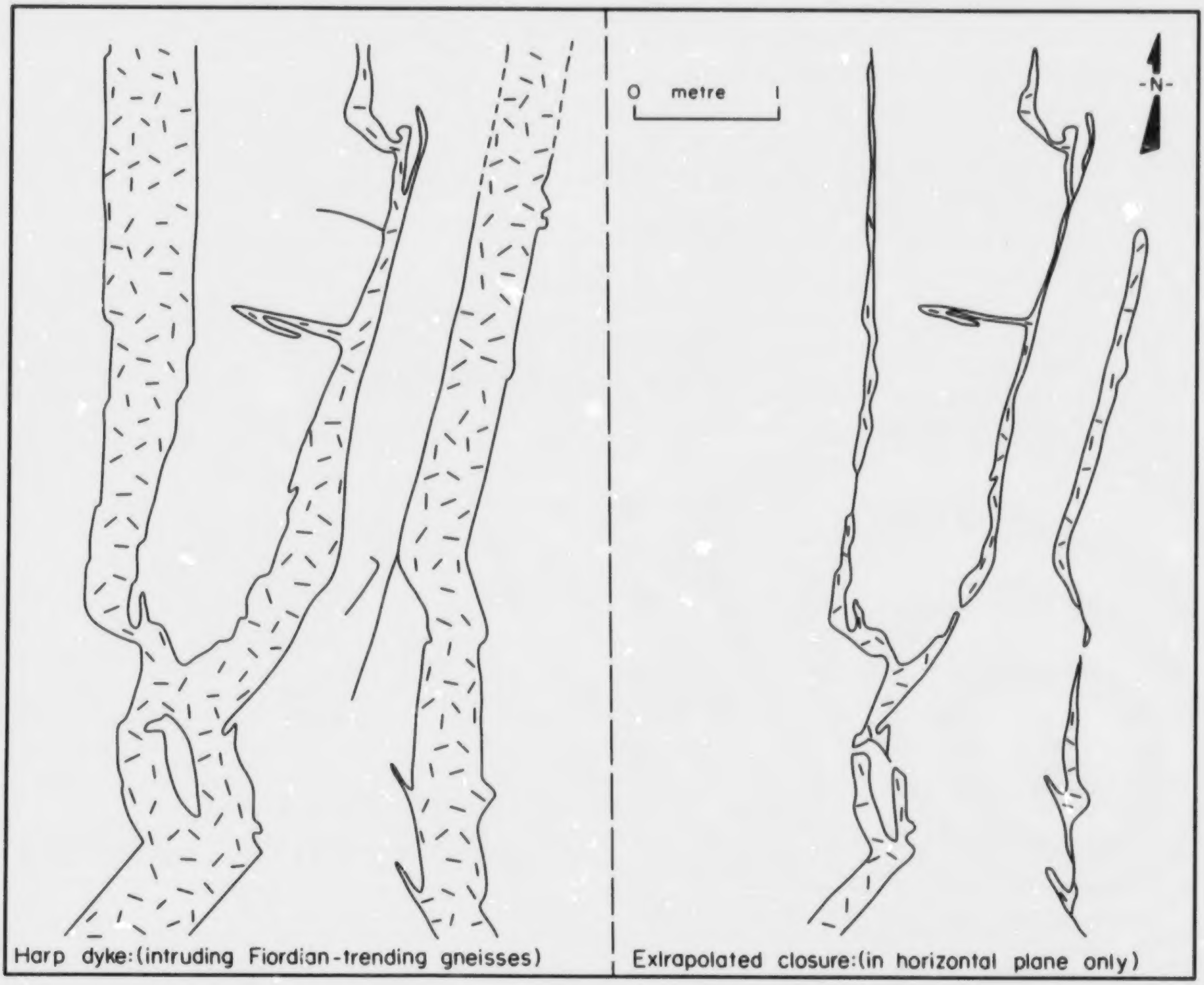
0 metre 1
N
Harp dyke: (intruding Fiordian-trending gneisses)
Extrapolated closure: (in horizontal plane only)

Figure 9. *Extrapolated closure of a Harp dyke as an indicator of the presence of vertical dilation components.*

ACKNOWLEDGMENTS

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GEOLOGY ALONG THE CENTRAL PART OF THE UGJOKTOK RIVER, (NTS 13N/5 AND PARTS OF 13M/8 AND 13N/6) LABRADOR

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ABSTRACT

The central part of the Ugjoktok River is underlain by early Precambrian polydeformed gneisses, granitoid and volcanic sedimentary rocks, which are intruded by Middle Proterozoic mafic to felsic plutonic massifs. Pyroxene- and amphibole-bearing tonalite orthogneiss of the Nain Province Archean Hopedale block has been metamorphosed to upper amphibolite–granulite facies during the Archean and/or Proterozoic orogenic events. Garnet-, sillimanite-, and cordierite-bearing quartzofeldspathic paragneiss of suspected Aphebian age has been deformed and metamorphosed to upper amphibolite grade during the Hudsonian Orogeny. The paragneiss is in contact with the orthogneiss along a folded, 0.5-km-wide zone, consisting of intensely mylonitized rocks of the latter unit. Paragneiss is intruded by foliated garnet and hypersthene-bearing porphyritic granodiorite. Westward, paragneiss and porphyritic granodiorite are interlayered with a series of amphibolite- to granulite-grade plutonic and gneissic rocks of the Churchill Province. These rocks have been intensely deformed and mylonitized during the Hudsonian Orogeny, to form a sequence of northerly trending straight belts.

Several kilometres east of the paragneiss–orthogneiss boundary, deformed Aphebian mafic to intermediate volcanic and sedimentary rocks (containing clasts of both the volcanic rocks and the basement gneisses), form the Ingrid group. Rocks of the Ingrid group have been folded and flattened during the Hudsonian Orogeny, and are in part thrust eastward over basement orthogneiss and in part lie unconformably upon it.

Fresh, igneous-textured leucogabbro, gabbronorite and anorthosite, intruded by porphyritic to equigranular, igneous-textured olivine-bearing monzonite, quartz monzonite and syenite, are found in both the Harp Lake and Nain igneous complexes. A locally rapakivi-textured variant of porphyritic olivine-bearing monzonite is found only within the Harp Lake complex. These undeformed, Paleohelikian rocks form extensive massifs that intrude orthogneiss, paragneiss and straight belts of granitoid gneiss along sharp, well-defined contacts. Although reported to the north of the Ugjoktok River area, no extensive, widespread metamorphic aureoles were defined in this study in the basement gneisses surrounding the massifs. A number of late Paleohelikian?, subhorizontal, olivine-bearing pyroxene gabbro to pyroxene diabase sills, both intrude and lie upon rocks of the Harp Lake complex as well as the basement gneisses and granitoids. Associated with these are hypabyssal or volcanic fragmental flow rocks. These sills are undeformed and exhibit pristine igneous textures. Neohelikian peralkaline granite and quartz–feldspar porphyry of the Flowers River Igneous Suite comprise the youngest rocks found in the area. Dykes and small bodies of peralkaline granite and quartz–feldspar porphyry, the latter probably as xenoliths or small roof pendants, occur within felsic granitoid rocks of the Nain igneous complex.

INTRODUCTION

This report discusses the results of a fill-in mapping project undertaken in a small area that was omitted earlier, during the course of three other regional mapping projects. The study area encompasses NTS sheet 13N/5 and the northern parts of sheets 13M/8 and 13N/6, which are located along the Ugjoktok River, approximately 255 km north-northwest of Goose Bay, Labrador (Figure 1). Final publication from this mapping project is intended to be at 1:100,000 scale. West and south of this area, a major regional study on the Harp Lake complex was carried out by Emslie (1980). Hill (1982) mapped the Flowers River and Nain igneous complexes to the north and to the east, Korstgård and Ermanovics (1984), Grant *et al.* (1983), Knight *et al.* (1981), Taylor (1979), Ermanovics and Raudsepp (1979), Ermanovics (1980), Ermanovics and Korstgård (1981) and Ermanovics *et al.* (1982) have reported on rocks of the Archean Hopedale block. Mapping of the intervening area had the additional objective of integrating these earlier surveys. The field component was carried out from June 20th to August 17th, and involved ground traversing, helicopter-borne reconnaissance as well as limited boat work. In addition, a geochronological program carried out under the direction of Urs Scharer at the Université du Québec à Montréal, resulted in the collection of seven samples from key lithostratigraphic units, to determine U–Pb zircon dates.

Regional Geology

The central Ugjoktok River area straddles the boundary between the Nain and Churchill provinces (Figure 1), and is underlain by a crystalline terrain consisting of

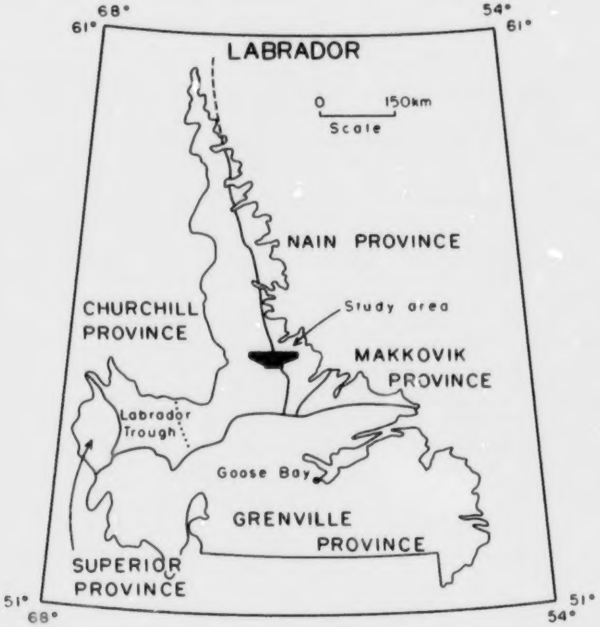


Figure 1. *Location of central Ugjoktok River map area (structural province divisions after Taylor, 1971; Wardle, 1983; and modified from Smyth and Greene, 1976).*

approximately sixty percent igneous rocks and forty percent amphibolite- to granulite-grade gneissic rocks. Beyond the map area to the northwest, are Proterozoic amphibolite- to granulite-grade migmatitic and agmatitic gneisses of the central core of the Churchill Province. Rocks in the northeastern part of the map area are continuous with the Paleohelikian pyroxene–amphibole granites and gabbroid plutons of the Nain igneous complex, which range in age between 1200 Ma and 1400 Ma (Hill, 1982), as well as the Neohelikian peralkaline volcanics and granite of the Flowers River Igneous Suite that are about 1270 Ma old (Hill, 1982). To the east and southeast, the Archean Hopedale block consists of amphibolite to granulite facies, granitic to tonalitic gneisses, migmatites, granitic intrusive rocks, greenstone belts and amphibolite. Extending to the west, southwest and south of the Ugjoktok River area, for up to 120 km, are granitic, anorthositic and gabbroic rocks of the Harp Lake complex, dated at 1450 ± 25 Ma (Emslie, 1980).

Paleohelikian and Neohelikian igneous rocks surrounding the map area are essentially undeformed and occur as large, roughly circular massifs. Proterozoic gneissic and granitic rocks of the Churchill Province are polydeformed, and have a predominantly northerly to north-northwesterly trend. The Archean Hopedale block lithologies are also polydeformed, exhibiting several trends, the most predominant being northeasterly.

General Geology of the Ugjoktok River Area

The map area is divided into twenty-one lithological types (Figure 2). The oldest rock present is amphibolite- to granulite-grade tonalite orthogneiss (subunit 1a), which is tentatively correlated with the Archean Maggo gneiss of Ermanovics *et al.* (1982). The tonalite gneiss is predominantly exposed in a block, north and east of the Ugjoktok River on map sheets 13N/5 and 13N/6 (Figure 2), where it can be traced into continuity with the Maggo gneiss to the east. Along its northwestern boundary, the tonalite gneiss is in contact with garnet paragneiss of Unit 3. The contact is marked by mylonite approximately 250- to 500-m wide, and consists of mainly tectonized tonalite orthogneiss (subunit 1b). Both the mylonite zone and the southwestern boundary of the tonalite gneiss are truncated by intrusive rocks of the Harp Lake complex.

Foliated pyroxene–hornblende–biotite granite (Unit 2) intrude the tonalite gneiss in the northern part of NTS 13N/6, but was not examined in detail.

Sillimanite and hypersthene-rich garnet paragneiss (Unit 3) correlate with an extensive unit of quartz, feldspar, garnet mylonitic gneiss located along the Nain–Churchill Province boundary in northern Labrador (Ryan *et al.*, 1983), and is known as the Tasiuyak gneiss (Wardle, 1983, 1984; Korstgård *et al.*, 1987). It underlies an 18-km-wide north–south-trending belt along the northern parts of 13N/5 and 13M/8 map sheets, and covers roughly the same area as the tonalite gneiss. South of Sarah Lake (Figure 2), several deformed quartz diorite to dioritic bodies (Unit 4) intrude the garnet paragneiss, possibly in the core of folds.

Throughout its area of exposure, garnet paragneiss is interlayered with garnet and hypersthene-bearing porphyritic granodiorite (Unit 5). Granodiorite is weakly to moderately foliated, intrusive into the garnet paragneiss, and in zones of intense local deformation, forms augen gneiss. Deformed plutonic and gneissic rocks of generally granitic, syenitic and dioritic compositions occur as relatively narrow north–south-trending bands (Units 6 to 12), in the northwestern part of map sheet 13M/8 (Figure 2). These rocks are continuous with those of the Churchill Province granitoid–gneiss belt of Hill (1982) in the adjoining Flowers River area to the north. Units 3 to 12, supposedly of Aphebian age and associated with some possible Archean component(s), all terminate against later intrusive rocks belonging to the Nain and Harp Lake complexes.

A supracrustal sequence of mafic to intermediate volcanic and derived sedimentary rocks, termed the Ingrid group (Unit 14) by Ermanovics and Korstgård (1981), unconformably overlies tonalite orthogneiss along the eastern edge of map sheet 13N/5. Although the group was originally subdivided into five units (Ermanovics and Korstgård, 1981; Knight *et al.*, 1981) for simplicity, the upper three have been combined and an additional one added. (It is noted here that these rocks were not remapped during this study). Collectively, the rocks of the group are overturned to the east, are presumed by Ermanovics and Korstgård (1981) to be Aphebian in age and lie unconformably upon the underlying Archean gneisses.

The Ingrid group is abruptly truncated at its northern end by a large igneous massif belonging to the Nain igneous complex, which underlies a deeply incised, barren highland plateau dominating the northeastern part of the map area (Figure 2). Leucogabbro and gabbronorite (subunit 15a) as well as subordinate anorthosite (subunit 15b) are intruded by later, areally dominant monzonite, quartz monzonite and syenite (Unit 16). These rocks have a well-defined, intrusive contact with tonalite gneiss of subunit 1a, but no intense, widespread metamorphic aureole was recognized. Rocks of both the massif and the Nain igneous intrusive bodies are continuous in the Flowers River area to the north, which yield Paleohelikian ages (Hill, 1982) of emplacement.

The southwestern and western parts of the map area are underlain solely by rocks of the Harp Lake complex, which forms a barren highland igneous massif, similar to that in the northeastern part of the area. It contains leucogabbro–gabbronorite (subunit 17a) and anorthosite (subunit 17b), intruded by monzonite to quartz monzonite (Unit 18). Rocks of this massif also intrude older lithologies including Units 1 through 13 (Figure 2), but show little evidence of contact metamorphism away from the contact. The Harp Lake complex was emplaced during the Paleohelikian (Emslie, 1980).

Unique within the Ugjoktok River map area is either a single or series of subhorizontal to horizontal pyroxene gabbro–diabase sill(s) (subunit 19c), which intrude(s) garnet paragneiss (Unit 3), quartz diorite (Unit 4), garnetiferous porphyritic granodiorite (Unit 5) and Harp Lake complex monzonite to quartz monzonite rocks (Unit 18). Fragmental (hypabyssal?) mafic flow rocks (subunits 19a and 19b) are closely associated, typically along the margins of the sill(s) (Figure 2), and contain some clasts from lithologies unrecognized within the map area. The Unit 19 gabbroic–diabasic rocks exhibit fresh intrusive or volcanic igneous textures, are undeformed and therefore inferred to be Paleohelikian.

Peralkaline granite (Unit 21) of the Neohelikian Flowers River Igneous Suite of Hill (1982), occurs in two places within the northeastern Nain igneous complex massif. It is closely associated with gabbronorite at the easternmost locality, and contains angular fragments of quartz–feldspar porphyry of Unit 20 at the locality north of the Ingrid group. The quartz–feldspar porphyry (Unit 20) is suspected of having peralkaline chemistry, and therefore, has been also assigned to the Flowers River Igneous Suite.

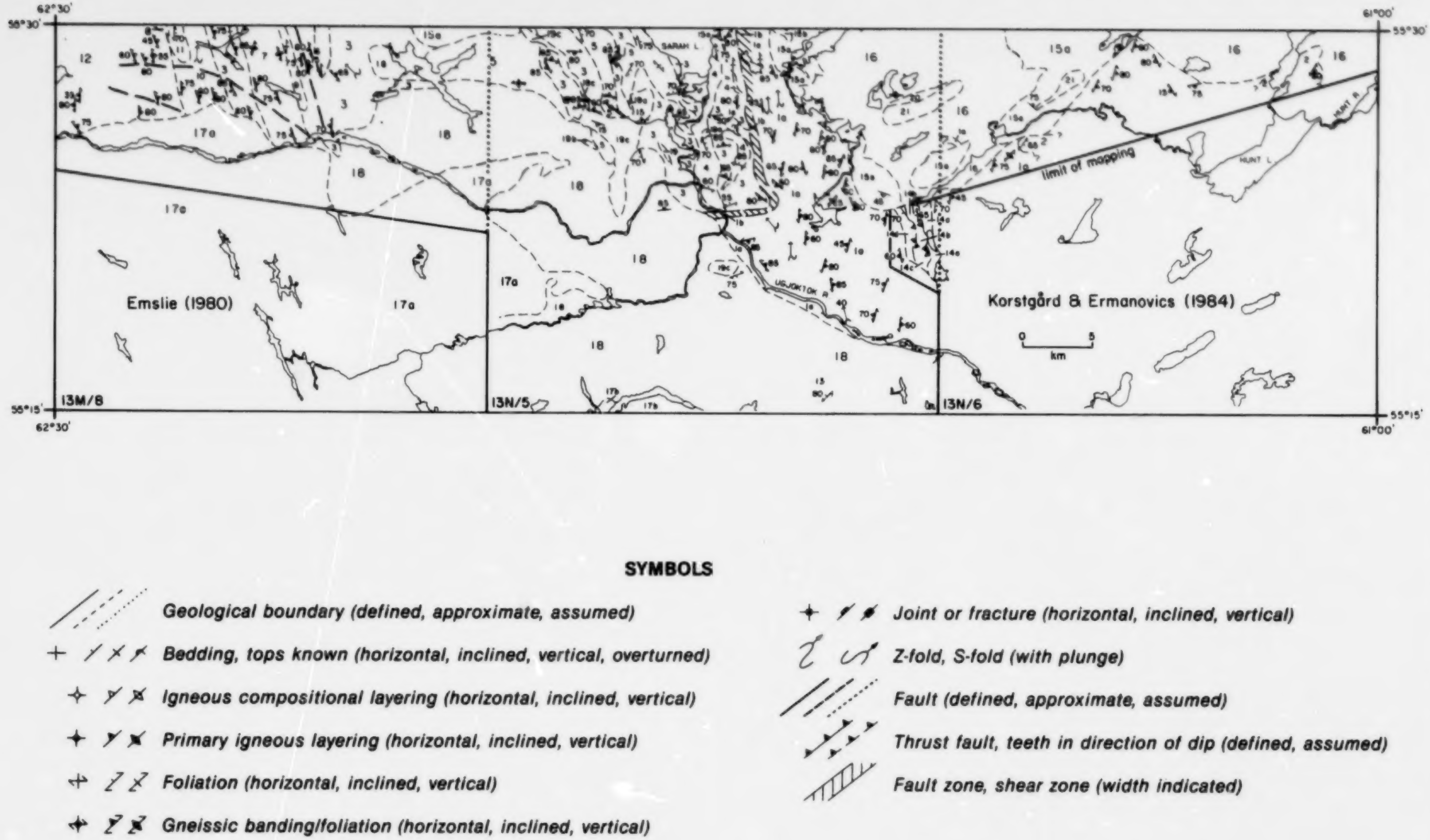
Gneissic and deformed granitoid rocks within the Ugjoktok River map area have undergone two known orogenic events. Tonalite orthogneiss (Unit 1) has been subjected to an Archean orogenic event, and at least north of the Ugjoktok River, has been reworked during the Proterozoic Hudsonian event. Both garnet paragneiss (Unit 3) and garnetiferous porphyritic granodiorite (Unit 5) have undergone Hudsonian deformation. In addition, the rocks of Units 6 through 13 have also been tectonized during the Hudsonian Orogeny but may contain Archean remnants affected by pre-Hudsonian tectonism.

Tonalite Orthogneiss (Unit 1)

Tonalite orthogneiss (subunit 1a) is well banded, but may be locally migmatitic or agmatitic. Banding may be regular and planar, or irregular and contorted (Plate 1), and consist of leucosome mobilizate layers ranging up to 5 cm in width, separating much wider (up to tens of metres) melanosome restite layers. Tonalite orthogneiss weathers grey to greenish-grey, is grey–green on fresh surface and consists of quartz, K-feldspar, plagioclase, orthopyroxene, green amphibole, brown biotite ± garnet ± magnetite ± chlorite ± epidote. The rock exhibits a fine- to medium-grained equigranular, polygonal, granoblastic texture and has a grain size up to 2 or 3 mm. Garnet occurs as fine-grained crystals within the matrix. The gneiss contains amphibolite bands up to tens of metres in width and is intruded by basic dykes, some of which are layer parallel and have been metamorphosed and recrystallized, others of which are fresh and cross-cut the gneissic banding. The cross-cutting dykes, which are generally narrow (< 1 m) and trend northeasterly, may belong to the Harp dyke swarm of Emslie (1980).

Plate 1. *Irregular, contorted banding in tonalite orthogneiss belonging to Unit 1.*

Mylonitic tonalite orthogneiss (subunit 1b), (Figure 2) is extremely fine-grained, recrystallized and dissected by numerous chlorite-filled fractures. In addition, it is locally brecciated and/or strewn with rounded feldspar porphyroclasts, up to 0.5 cm in diameter. On the weathered surface it is rusty pink and on the fresh surface it is greenish-red. The mylonitization may be due to a single fault, or may represent the root of a higher level fault zone. The subunit 1b orthogneiss zone is itself folded, and therefore probably represents a relatively early structure.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13M/8
13N/5
13N/6
Emslie (1980)
Korstgård & Ermanovics (1984)
limit of mapping
SARAH L.
HUNT L.
HUNT R.
UGJOKTOK R.
0
5
km
12
3
15a
16
17a
18
SYMBOLS
Geological boundary (defined, approximate, assumed)
Bedding, tops known (horizontal, inclined, vertical, overturned)
Igneous compositional layering (horizontal, inclined, vertical)
Primary igneous layering (horizontal, inclined, vertical)
Foliation (horizontal, inclined, vertical)
Gneissic banding/foliation (horizontal, inclined, vertical)
Joint or fracture (horizontal, inclined, vertical)
Z-fold, S-fold (with plunge)
Fault (defined, approximate, assumed)
Thrust fault, teeth in direction of dip (defined, assumed)
Fault zone, shear zone (width indicated)

Figure 2. *Geology along the central part of the Ugjoktok River, Labrador.*

LEGEND

NEOHELIKIAN

FLOWERS RIVER IGNEOUS SUITE

21 *Peralkaline granite*

20 *Quartz, feldspar porphyry*

PALEOHELIKIAN

19 19a, *Fragmental flow rock (consists of angular fragments of a variety of lithologies in a matrix of amygdule-rich hypabyssal mafic rock:* 19b, *fragmental flow rock similar to subunit 19a, but containing large quantities of graphite;* 19c, *pyroxene gabbro to pyroxene diabase*

HARP LAKE COMPLEX

18 *Monzonite, quartz monzonite (commonly rapakivi textured)*

17 17a, *Leucogabbro, gabbronorite;* 17b, *anorthosite*

NAIN IGNEOUS COMPLEX

16 *Monzonite, quartz monzonite, syenite*

15 15a, *Leucogabbro, gabbronorite;* 15b, *anorthosite*

APHEBIAN

INGRID GROUP

14 14a, *Mafic to intermediate lavas, mafic conglomerate, sandstone, siltstone;* 14b, *mafic to intermediate lavas, porphyritic basalt, mafic conglomerate sandstone, siltstone;* 14c, *mafic volcanic conglomerate, felsic polymictic conglomerate, sandstone siltstone, grit, mudstone;* 14d, *mylonitized and tectonically intercalated rocks of Unit 1 and Ingrid group*

APHEBIAN and/or ARCHEAN

13 *Foliated equigranular granite*

12 *Granitic orthogneiss*

11 *Foliated hornblende, pyroxene granite*

10 *Foliated biotite, amphibole diorite to monzodiorite*

9 *Foliated quartz syenite*

8 *Amphibolite, includes minor bands of diorite, tonalite gneiss and schist*

7 *Granitic gneiss intermixed with tonalite gneiss and quartz syenite*

6 *Porphyritic granitic gneiss*

5 *Garnet, hypersthene-foliated porphyritic granodiorite to augen gneiss*

4 *Quartz diorite, diorite*

3 *Garnet paragneiss with or without sillimanite and hypersthene*

ARCHEAN

2 *Foliated hornblende, biotite granite*

1 1a, *Tonalite orthogneiss;* 1b, *mylonitic tonalite orthogneiss*

The type and homogeniety of mineralogy over large expanses of the tonalite gneiss, suggest a dioritic to quartz dioritic igneous protolith. The gneiss has had a complicated tectonometamorphic history and has attained upper amphibolite to granulite facies conditions, and hence may be as old as Archean.

Hornblende–Biotite Granite (Unit 2)

Rocks belonging to the two limited exposures of granite on map sheet 13N/6, are moderately to strongly foliated, having patches of gneissic texture developed in the easternmost exposure. They weather buff-white to pink, are pink–white on fresh surface and comprise quartz, K-feldspar, plagioclase, biotite, amphibole and orthopyroxene. The rocks of Unit 2 appear to intrude the tonalite orthogneiss of Unit 1, however, both these rock types have undergone at least one common deformational event.

Garnet Paragneiss (Unit 3)

Quartzofeldspathic paragneiss (Plate 2) weathers white, is grey–green to grey on fresh surface and exhibits medium to coarse polygonal granoblastic texture. It is well banded, having cordierite, biotite, minor muscovite, hypersthene?, fine-grained subhedral to euhedral garnet, magnetite and/or ilmenite and commonly associated with coarse prismatic sillimanite confined to melanocratic restite bands. These bands vary in width from 2 to 10 cm. Leucocratic mobilizate bands contain quartz, feldspar and large (up to 2 cm) euhedral to subhedral garnet porphyroblasts (Plate 2). In general, leucocratic bands are much wider than their melanocratic counterparts, quite commonly occurring on a scale of one to two metres. Banding in the gneiss may be regular or contorted, but generally represents polyphase deformation. Mushroom structures and chevron folds, as well as transposition structures such as hook folds, are common throughout the unit.

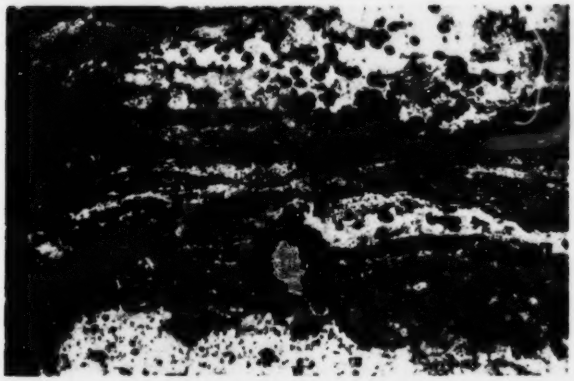

Plate 2. *Garnet porphyroblasts in quartzofeldspathic garnet paragneiss of Unit 3.*

The presence of abundant sillimanite suggests that the protolith was a pelitic sediment. Both the protolith and deformation within the gneiss are considered to be Proterozoic, the latter being due to the Trans-Hudson Orogen. Pending the results of a comprehensive geochronological program, for which samples from this unit were collected, the possibility of parts or all of the paragneiss being as old as Archean is not yet discounted.

Quartz Diorite–Diorite (Unit 4)

The quartz diorite to dioritic rocks of Unit 4 consist of medium-grained, recrystallized quartz, plagioclase, hypersthene, amphibole, biotite and magnetite and/or ilmenite. In the cores of the two large bodies, south of Sarah Lake (Figure 2), relict medium- to coarse-grained igneous textures prevail, whereas polygonal, granoblastic textures are typical in the small body southeast of the lake, as well as in the margins of the larger bodies. These rocks are intrusive into Unit 3 paragneiss, however, both their absolute age relationship and tectonic history with respect to the gneiss are uncertain.

Garnet–Hypersthene Porphyritic Granodiorite (Unit 5)

As with garnet paragneiss, porphyritic granodiorite weathers white and is buff-white or white on fresh surface. It exhibits a characteristically coarse porphyritic texture (Plate 3) containing euhedral to subhedral, elongate K-feldspar crystals up to 5 cm long in a medium-grained quartz, plagioclase, orthopyroxene and amphibole matrix. Coarse subhedral to rounded garnet and/or orthopyroxene? porphyroblasts up to 1.5 cm in diameter are ubiquitous throughout the rocks of this unit (Plate 4). Porphyritic granodiorite is strongly foliated to gneissic and the feldspar phenocrysts show marked tectonic alignment. Locally, the rocks of Unit 5 are migmatitic and are almost everywhere intruded by criss-crossing dykes and veins of felsic aphanitic material.

Plate 3. *Coarse porphyritic texture in foliated garnet, hypersthene granodiorite of Unit 5; note contact of granodiorite with Unit 3 paragneiss at upper right.*

Porphyritic granodiorite has been metamorphosed under conditions of amphibolite to granulite facies. It is itself deformed, but also intrudes previously deformed Unit 3 paragneiss (Plates 3 and 5), indicating either a late syntectonic

Plate 4. *Closeup of garnet porphyroblasts within Unit 5 garnet, hypersthene granodiorite.*

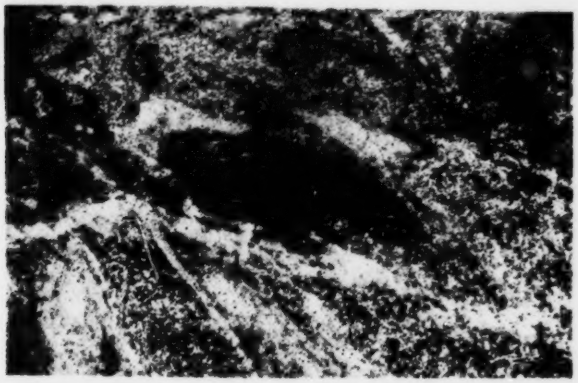

Plate 5. *Xenolith of deformed Unit 3 paragneiss surrounded by Unit 5 granodiorite.*

origin with respect to deformation in the paragneiss, or posttectonic intrusion, if subsequently deformed.

Churchill Province Straight Granitoid–Gneiss Belts (Units 6 to 12)

These lithologies include porphyritic granite gneiss (Unit 6), amphibolite (Unit 8), quartz syenite (Unit 9), diorite–monzodiorite (Unit 10) and granitic orthogneiss (Unit 12), as well as several other minor granitoid and gneiss units. Two features link all these units: 1) the upper amphibolite to granulite facies metamorphic mineral assemblages, and 2) the occurrence of variously developed mylonitic fabrics. The mineralogy of the rocks within these units, comprises quartz, K-feldspar, plagioclase, orthopyroxene, clinopyroxene, amphibole, biotite, garnet, sillimanite and chlorite. Orthopyroxene (hypersthene?) is present in approximately seventy percent of the felsic gneissic and granitic rocks. Chlorite is a minor retrograde alteration product of amphibole, garnet and pyroxene.

Fifteen of seventeen samples examined in thin section showed moderately to strongly developed mylonitic textures, including highly strained and intensely recrystallized quartz and feldspar. This sequence of Churchill Province rocks is monotonous and was examined in detail by Hill (1982).

Foliated Equigranular Granite (Unit 13)

Two small outcrops of foliated equigranular granite occur south of the Ugjoktok River as large xenoliths or possible roof pendants in monzonitic to quartz monzonitic rocks of the Harp Lake complex. This granite consists of quartz, microcline, orthoclase, amphibole and magnetite and/or ilmenite. Both its origin and absolute age are unknown.

Ingrid Group (Unit 14)

Detailed descriptions of rocks belonging to this unit have been given by Knight *et al.* (1981) and Ermanovics and Korstgård (1981), from which the following description is summarized. An overturned synform along the eastern edge of the group contains mafic to intermediate lavas, mafic volcanic conglomerate, sandstone and siltstone (subunit 14a). The lavas comprise mainly massive, or rarely, pillowed subaerial basalts and andesites, characterized by a fine-grained matrix of chlorite, epidote and actinolitic amphibole. Individual flows are not recognizable. The associated conglomerate contain clasts, which are derived from these mafic lavas.

The central part of the group is dominated by an overturned antiform, where, in the northern part, mafic to intermediate lavas, porphyritic basalt, mafic volcanic conglomerate, sandstone and siltstone are exposed (subunit 14b). The porphyritic basalt is characterized by irregularly distributed plagioclase phenocrysts up to 2 cm long, whereas mafic to intermediate lavas are identical to those in subunit 14a. Mafic conglomerate in this unit contain clasts derived from the porphyritic basalt, and include unsorted, poorly bedded rocks and angular fragments comprising basalt, gabbro and shale, as well as cobbles and boulders of the underlying Unit 1 orthogneiss.

Both the synform and antiform are overturned toward the east and bounded in part along their limbs by west-dipping thrust faults. The south-central section of the group contains mafic volcanic conglomerate, felsic polymictic conglomerate, sandstone, siltstone and grit and mudstone of the uppermost formation (subunit 14c). Clasts within this unit are also derived from volcanic and sedimentary rocks as well as basement gneiss. They include granodiorite, perthitic granite, diabase, sandstone, siltstone, shale, quartzite and limestone.

Subunit 14d along the western border of the Ingrid group, comprise mylonitized and tectonically intercalated rocks of Unit 1 basement gneiss and the group itself. The unit forms a 1-km-wide band along the edge of the group, and the contact with orthogneiss, *sensu stricto*, is gradational and obscured by mylonite. In this report, this unit is included within the group, but Ermanovics and Korstgård (1981) and Knight *et al.* (1981), mapped it as a separate lithological division outside of the group.

Conglomerates within the group are generally poorly sorted and the graded beds have indistinct gradational boundaries. Soft sediment slump structures, large isolated clasts in finer grained beds, and coarse unsorted detritus in beds of unsorted felsic sandstone are also common. Overall, these features suggest debris slide deposits indicative of an unstable depositional environment. Volcanic rocks intercalated with coarse polymictic detritus derived from both basement and underlying lavas, may be also indicative of volcanism in a tectonically unstable environment. The Ingrid group has been metamorphosed to the greenschist facies during the Hudsonian Orogeny.

Nain Igneous Complex (Units 15 and 16)

The plutonic rocks of this complex generally fall into two compositional groupings, namely 1) leucogabbroic to gabbronoritic and anorthositic rocks and, 2) monzonitic and syenitic rocks. Leucogabbro and gabbronorite (subunit 15a) are massive and exhibit well-developed, coarse-grained equigranular, ophitic to subophitic textures. These rocks weather grey–green and are grey or grey–white on fresh surface and have grain sizes up to 4 cm. They contain plagioclase, hypersthene and/or clinopyroxene, magnetite, ilmenite, apatite and locally olivine.

Anorthosite (subunit 15b), which is closely associated, spatially, with the leucogabbro and gabbronorite, also exhibits coarse-grained, equigranular igneous texture. Where examined, it comprises approximately eighty-five to ninety percent plagioclase with minor hypersthene and opaque minerals.

Unit 16 includes monzonite, quartz monzonite and syenite. The rocks are buff-white to pink and commonly exhibit both a rusty and red-coloured weathering. Exposed surfaces are deeply weathered, friable and associated with talus piles of grus. This latter characteristic is marked in some places, where entire outcroppings of rock have weathered *in situ* to give a crust consisting of piles of equant feldspar crystals. The rocks comprise quartz, perthite, plagioclase, ortho- and clinopyroxene, amphibole, olivine, deep brown biotite, iddingsite, apatite, chlorite, epidote, magnetite, ilmenite and pyrite. The syenite is characterized by a quartz-free, olivine-bearing assemblage. Feldspars exhibit string, vein and braid perthites and both micropegmatitic and myrmekitic textures are common. Coarse-grained, equigranular, plutonic igneous textures predominate in these rocks, and porphyritic textures are rare or absent.

Harp Lake Complex (Units 17 and 18)

The Harp Lake complex rocks are divided into two groupings, namely 1) leucogabbro, gabbronorite and anorthosite, and 2) monzonite and quartz monzonite. Leucogabbro and gabbronorite (subunit 17a) are massive, and like their counterparts in the Nain igneous complex, exhibit fresh, coarse-grained, equigranular, igneous textures and have the same colour and mineralogy.

Subunit 17b anorthosite is identical in character and mineralogy to subunit 15b anorthosite. Monzonite and quartz monzonite of Unit 18 differ mainly from that of Unit 16 in having a porphyritic texture, commonly with locally developed, rapakivi-textured perthite cores mantled by plagioclase rims (Plates 6 and 7). Plutonic rocks of Unit 18 comprise quartz, perthite, microcline, plagioclase, orthopyroxene, clinopyroxene, olivine, amphibole, minor brown biotite, iddingsite, minor chlorite, magnetite and ilmenite. Locally, the presence of numerous granitoid and intensely metamorphosed gneissic xenoliths (Plate 8), suggests that Unit 18 plutonic rocks may represent a high level of intrusion.

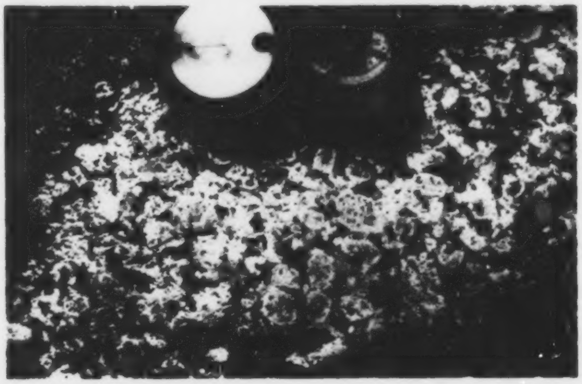

Plate 6. *Closeup of rapakivi texture in Harp Lake Complex quartz monzonite belonging to Unit 18.*

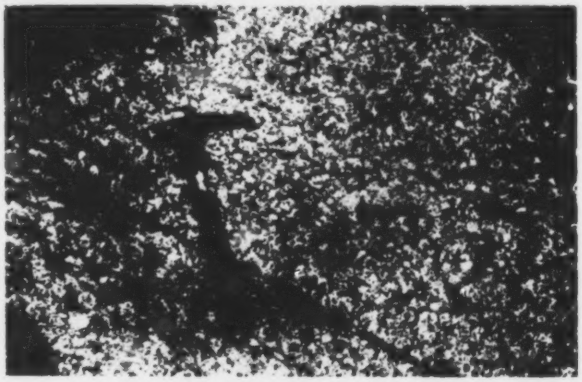

Plate 7. *Porphyritic, rapakivi-textured Harp Lake Complex quartz monzonite of Unit 18.*

Fragmental Flow Rock–Pyroxene Gabbro (Unit 19)

Subhorizontal to horizontal sills (Plate 9), southwest of Sarah Lake, contain mainly pyroxene gabbro to diabase, however, lesser amounts of hypabyssal or possibly volcanic flow rock occur along their margins.

Fragmental flow rock (subunit 19a) is indicative of high levels of intrusion. It consists of a variety of angular fragments (Plate 10), in a fine-grained mafic matrix. Fragments, which

Plate 8. *Granitoid and gneiss xenoliths in high-level intrusive Unit 18 quartz monzonitic rocks belonging to the Harp Lake Complex.*

Plate 9. *Subhorizontal sill of subunit 19c pyroxene gabbro resting upon altered quartz monzonitic rocks belonging to Unit 18 of the Harp Lake Complex; note weathering and breakdown of rocks along the contact.*

Plate 10. *Angular fragments of metaquartzite, gneiss and granitoids in a fine-grained mafic matrix; exposure is subunit 19a fragmental flow rock.*

range up to 0.5 m in diameter, consist of metaquartzite, felsic granitoid rocks and gneiss, are unsorted, and have a chaotic distribution suggestive of magmatic stoping. Parts of this unit are especially rich in fragments, almost to the exclusion of matrix material (Plate 11). Also present within the matrix are quartz and calcite-filled amygdules up to 1 cm in diameter. These are sparsely distributed and undeformed, with quartz commonly present as euhedral vug-filling crystals. The matrix of subunit 19a rocks is massive, exhibits a fresh pilotaxitic igneous texture, and consists of plagioclase laths in a groundmass of intersertal clinopyroxene, for the most part altered to chlorite and sericite. Locally, graphite and/or specular hematite and radiating clusters of zeolites? are also present. A chilled marginal contact (Plate 12) is locally exposed between fragmental flow rock (subunit 19a) and the country rocks.

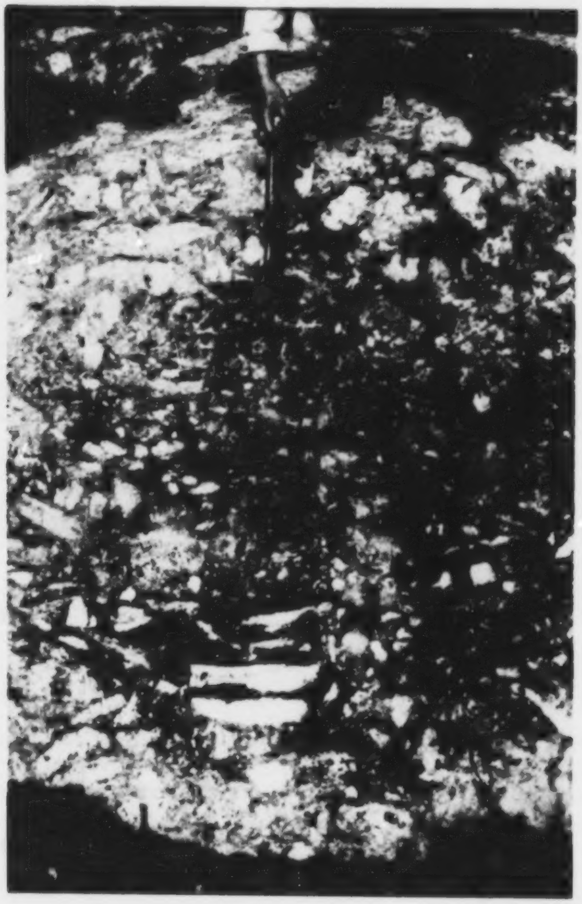

Plate 11. *Exposure is same area as that of Plate 10, showing close concentration of abundant fragments.*

A limited area of outcrop north of the western part of the Ugjoktok River consists of fragmental flow rock (subunit 19b) similar to that of subunit 19a, but contains large quantities

Plate 12. *Contact between fragmental flow rock of subunit 19a and quartz dioritic rock of Unit 4; note chilled margin of former against latter.*

of graphite and minor amounts of chalcopyrite and pyrite. In this unit, amygdules are not observed and the matrix texture may be in part detrital sedimentary. Therefore, although this rock is assigned to Unit 19, it may in actual fact be a raft of some earlier lithology, possibly correlative with the Ingrid group.

Pyroxene gabbro to pyroxene diabase (subunit 19c) is fresh, exhibits a massive igneous texture, and at several localities near the base of the sills has well-developed igneous layering. In general, there is a fining of grain size toward the lower margin of the sills, with the gabbroic textures at higher levels, giving way to diabasic textures along a chilled marginal contact. The basal contact of one prominent sill (Plate 9) is marked by a 1.5-m-thick band of altered, iron-stained pyrite, pyrrhotite-rich, fine-grained rock, which is quite noticeable from a great distance, and can be mapped easily from the air. Subunit 19c gabbro–diabase contains plagioclase, orthopyroxene, clinopyroxene, olivine, red–brown biotite, magnetite and/or ilmenite and, where altered, minor chlorite, iddingsite, serpentine and calcite.

Flowers River Igneous Suite (Units 20 and 21)

Quartz-feldspar porphyry (Unit 20) occurs in a narrow, 1.5-km band northeast of the Ugjoktok River. It weathers to a rusty-buff colour and is reddish-green on fresh surface. The rock is massive, fresh and contains mainly euhedral to subhedral K-feldspar and subordinate quartz phenocrysts up to 1 cm across, in a groundmass of fine-grained polygonal quartz, K-feldspar, plagioclase, clinopyroxene, amphibole, olivine? and magnetite and/or ilmenite. K-feldspar phenocrysts are zoned and clinopyroxene crystals are commonly embayed and resorbed. At present, it is considered that the quartz–feldspar porphyry is peralkaline, as the clinopyroxene is suspected to be aegirine. In addition, the porphyry is intruded by suspected peralkaline granite (Plate 13).

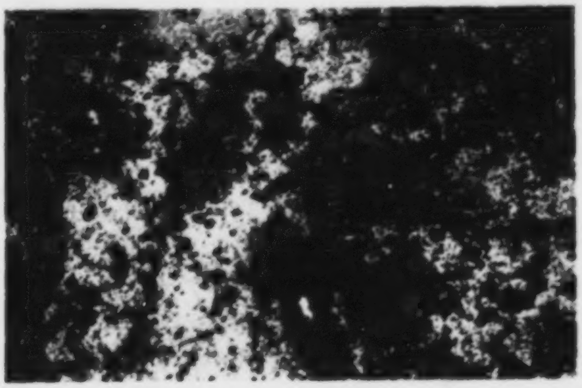

Plate 13. *Subangular xenoliths of Unit 20 quartz, feldspar porphyry in Unit 21 peralkaline granite.*

Peralkaline granite (Unit 21), closely associated, spatially, with the quartz–feldspar porphyry, is massive and fresh, exhibiting a coarse plutonic igneous texture. One suspected occurrence of this rock type is as a small elongate body in Nain igneous complex rocks north of the Ingrid group, however, it is more commonly found as narrow dykes at localities widespread throughout the complex. Peralkaline granite weathers buff-white or pink–white, is medium- to coarse-grained and contains quartz, perthite, plagioclase, aegirine and arfvedsonite and/or riebeckite.

Structure

Rocks within the Ugjoktok River map area are divided into three structural groups: 1) those that are undeformed, 2) those deformed during the Hudsonian Orogeny, and 3) those deformed during the Hudsonian and/or earlier (Archean) events. Apart from some localized shearing and mylonitization, igneous rocks belonging to the Nain igneous complex, Harp Lake complex and Flowers River Igneous Suite are essentially undeformed. No evidence was observed to suggest faulting of any great magnitude.

It is suspected that rocks of the Ingrid group are Aphebian and therefore, were deformed during the Hudsonian Orogeny. Knight *et al.* (1981) and Ermanovics and Korstgård (1981) interpret the regional structure of the group as resulting from an originally gently flexed anticline, subsequently folded and faulted along steep, west-dipping thrust faults. These would have most likely occurred by slippage along the limbs of overturned large-scale regional folds.

Basement gneisses deformed during the Hudsonian Orogeny are separated into two groups. Churchill Province gneisses of map sheet 13M/8, within the northwestern part of the area, are flattened and straightened, exhibiting an intense, regular, steep-dipping, north-trending gneissic banding and/or foliation (Figure 3b). These northwestern

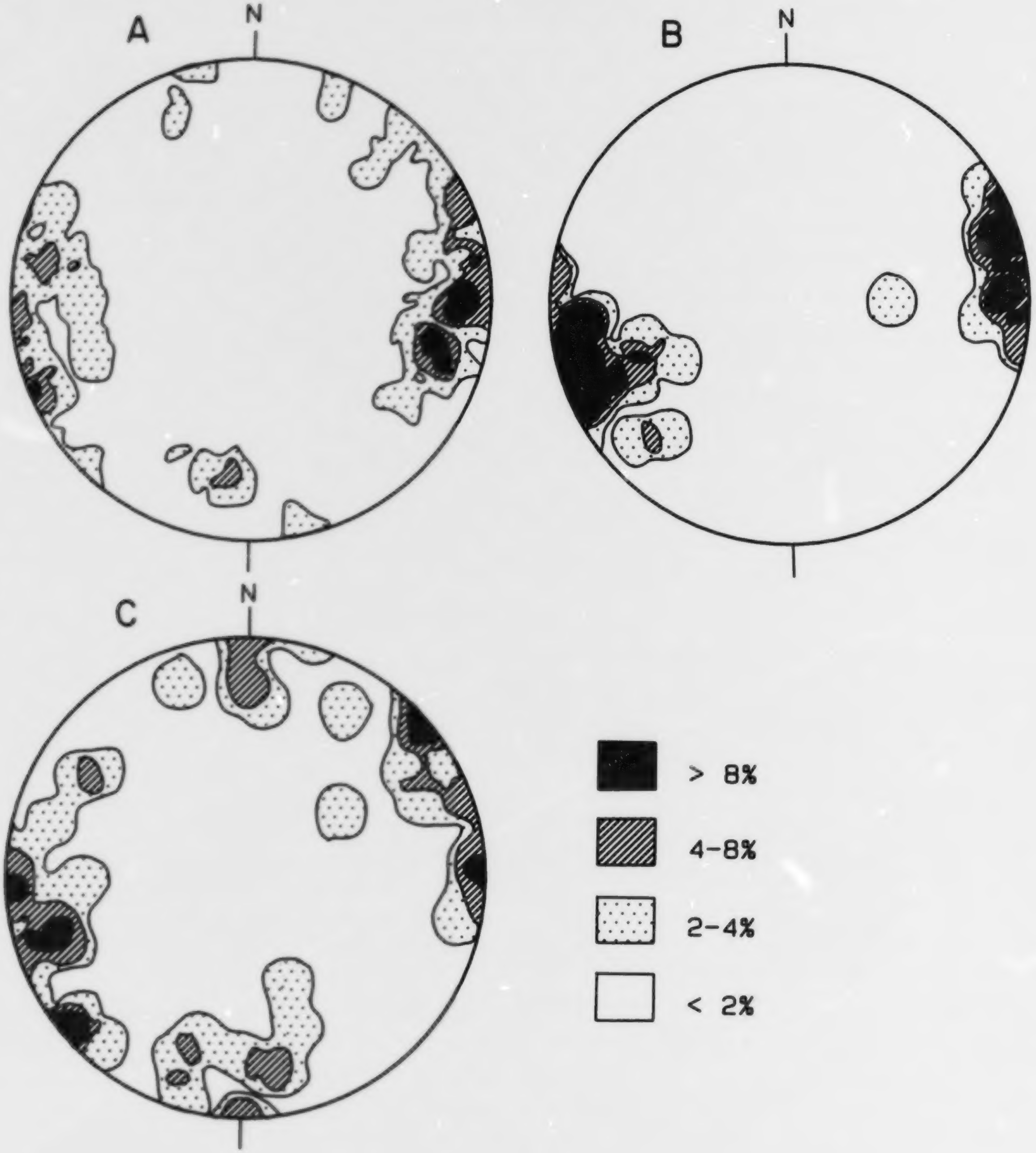

Figure 3. *Contoured stereographic plots of poles to gneissic banding–foliation planes for, a) the central gneisses; b) the northwestern gneisses; and c) the foliation planes for central granitoid rocks; plots are based on 80, 30 and 36 data points respectively; contour intervals are in percent of poles per one percent area of the stereonet.*

gneisses are polydeformed, and both mesoscopic and megascopic folds are closed, isoclinal structures, with the observed folds being flattened and occasionally recumbent. At one locality in the extreme northwestern part of the area, gneissic banding is nearly horizontal on the vertical face of an outcrop, and nearly vertical on the top of the same outcrop. It is possible that either two limbs of a fold are being observed, or that at this locality, an intense subvertical foliation overprints a flat-lying gneissic banding. For the amount of time spent examining the northwestern gneisses, it was not possible to determine a detailed breakdown into structural events and or domains.

Gneisses within the central part of the area exhibit greater structural diversity. As in the northwestern gneisses, gneissic banding and/or foliation in the central gneisses (Units 1 and 3) dip steeply east or west and trend north–south (Figure 3a), but is not as consistently flattened nor as tightly constrained to a north–south orientation. These characteristics are demonstrated by comparison of the stereoplots of Figures 3a and 3b. The pole maxima for the central gneisses are less intense than those for the northwestern gneisses. In addition, both pole maxima and pole girdles for the central gneisses are more dispersed, exhibiting a much wider variation in trend. Folding in the central gneisses is more open than in the northwestern gneisses. Regional, megascopic open to isoclinal F_2 folds plunge at moderate angles to the south or southeast, and refolded F_1 folds, such that the two become coaxial, with axial planes of both being vertical or subvertical. Gneissic banding and/or foliation were formed prior to, or during the of F_1 folding.

Deformed granitic rocks in the central part of the area exhibit foliation trends (Figure 3c) similar to the gneissic banding and/or foliation trends found within the surrounding gneisses. The major difference is the occurrence of a number of east–west trending, moderate to steep north- and south-dipping foliation planes, which are apparent on a stereoplot of the foliation data (Figure 3c).

Gneisses deformed during Hudsonian and/or earlier Archean events are confined to the southern and eastern parts of the map area. The tonalite orthogneiss (Unit 1) is believed to be of Archean age and to have primarily Archean structural features. West of the Ingrid group, and, according to Korstgärd and Ermanovics (1984), east of the group for several kilometres, Hudsonian reworking has imparted a northerly trend to the gneiss. Farther to the east, near Hunt Lake in the region examined by Ermanovics and others, east and northeast late Archean trends (Ermanovics *et al.*, 1982) predominate. According to Korstgärd and Ermanovics (1984), the eastern limit of Hudsonian deformation (and therefore the Churchill–Nain province boundary) is marked roughly by the Ingrid group and its nearby basement.

The (subunit 1b) mylonite zone may represent the root of a deep-seated thrust or reverse fault and constitute a more fundamental division between Proterozoic (Unit 3), and Archean (Unit 1) gneisses. A geochronological study, in progress, should define the Churchill–Nain province boundary more accurately in this area.

Several faults cut the northwestern gneisses and are marked by mylonite and deeply incised lineaments. Neither sense of movement nor age of these structures could be determined. Much of the northwestern gneiss terrain is mylonitic, having ribbon and strained quartz showing undulose extinction, fine-grained polygonal recrystallization and rotation features around rounded garnet; feldspar porphyroblasts are common. These Churchill Province gneisses have undergone associated extreme flattening.

Metamorphism

Apart from the local greenschist grade of late-stage deuteric alteration, rocks belonging to the Nain igneous complex, the Harp Lake complex, the Flowers River Igneous Suite and the fragmental flow rock and pyroxene and/or diabase of Unit 19, are essentially unmetamorphosed. The late-stage alteration is manifested by cross-cutting veinlets filled with chlorite, epidote, sericite and calcite and also by selective sericitization of plagioclase. Orthopyroxene and clinopyroxene are common mineral constituents of both gabbroid and granitoid plutons within these igneous complexes. Although these minerals are not in themselves of metamorphic origin, they indicate intrusion of the plutons under high-temperature anhydrous conditions. No obvious intense, widespread metamorphic aureole occurs about the igneous massifs, however, pyroxene exhibits retrograde alteration to amphibole at a number of localities in tonalite orthogneiss within several kilometres of the Nain igneous complex.

The occurrence of chlorite, epidote and actinolitic amphibole in rocks of the Ingrid group is indicative of greenschist facies metamorphism (Knight *et al.*, 1981; Ermanovics and Korstgärd, 1981).

All other gneissic and deformed granitoid rocks within the map area exhibit amphibolite to granulite facies mineral assemblages. The northwestern gneisses have an upper amphibolite–granulite Hudsonian metamorphic mineral assemblage containing orthopyroxene, clinopyroxene, garnet and biotite with or without sillimanite and cordierite. An amphibolite facies retrograde overprint of unknown age is indicated by the partial alteration of pyroxene to amphibole.

Both orthogneiss and paragneiss from the central and eastern parts of the map area contain the granulite facies assemblage of orthopyroxene, clinopyroxene, biotite and garnet. In addition, paragneiss contains sillimanite as a stable phase. A greenschist to amphibolite retrogression is also apparent in these rocks, indicated by the presence of secondary amphibole, chlorite, sericite and minor epidote. Until results from the geochronology sampling program are received, no attempt is made to date the metamorphism other than to state that the northwestern gneisses are primarily Hudsonian, that the central gneisses are Hudsonian and/or late Archean?, and that the eastern gneisses are primarily late Archean.

Economic Geology

Igneous rocks belonging to both the Nain and Harp Lake complexes contain abundant coarse magnetite and ilmenite. Although no layered concentrations of ilmenite were observed anywhere, rich accumulations of ilmenite are present within subunit 15a gabbronorite, in the northeastern part of the area. Crystals in excess of 2-cm length were observed at several localities. Subunit 17a gabbronorite is also rich in coarse ilmenite crystals. Both the Nain and Harp Lake complexes are singled out as possible exploration targets for chromium and platinum group transition elements.

Peralkaline granite (Unit 21) of the Flowers River Igneous Suite is enriched in Zr, Ce, La and other rare-earth elements. Several Zr-rich samples have been collected by Hill (1982) from Flowers River Igneous Suite rocks to the north. One locality within the peralkaline granite body north of the Ingrid group gave a total count scintillometer reading of 700 to 1000 cps.

Stringer chalcopyrite, pyrite, pyrrhotite mineralization, along with abundant graphite, is present within subunit 19b fragmental rocks. Disseminated sulphide mineralization also occurs at numerous localities within the northwestern Churchill gneisses. A number of rusty mineralized zones occur to the west of Sarah Lake. Of those examined, only disseminated pyrite and rusty hematitic weathering products were found. Some localities were exceedingly difficult to sample due to the intensely weathered nature of the rocks.

DISCUSSION

The geological nature of the Churchill–Nain boundary in the region of the central Ugjoktok River is still uncertain. In the absence of constraining geochronological data, only generalizations can be made. The northwestern gneisses are continuous northward along a wide belt of known Hudsonian deformation and are by definition, part of the Churchill Province. The northwestern gneiss terrain and associated deformation, can be traced continuously into the gneisses within the central part of the map area. Throughout both of these areas, rocks are characterized by north–south gneissic banding and/or foliation and fold trends. Across the subunit 1b mylonite zone, north–south trends persist and continue as far as 5 km east of the Ingrid group. From there eastward, easterly to northeasterly late Archean trends dominate. Until the significance of both the subunit 1b mylonite zone and the associated transition across it, from equivalents of the Tasiuyak gneiss to equivalents of the Maggo gneiss can be determined, it is suggested that the Churchill–Nain province boundary in the Ugjoktok River region remain as defined by Korstgård and Ermanovics (1984).

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STUDIES IN THE SEAL LAKE GROUP, CENTRAL LABRADOR, PART I: DESCRIPTION OF COPPER OCCURRENCES

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ABSTRACT

The Seal Lake Group, the youngest supracrustal sequence in the Labrador Central Mineral Belt, is a dominantly continental assemblage of red shale, siltstone, and quartzite, associated with basalt flows and numerous diabasic to gabbroic sills. The middle three formations and sills contain over 250 copper occurrences, ranging from malachite stains to significant chalcocite−bornite deposits. In general, native copper-bearing quartz−carbonate veins occur in red slate and/or strongly hematized basalt units, whereas, chalcocite and/or bornite stringers and quartz−carbonate veins occur in diabase, basalt, or grey to black shales and phyllite; the copper mineralogy being dependant upon the valency state of iron in the host rock. All occurrences appear to be epigenetic and are interpreted as the products of a single mineralizing event.

INTRODUCTION

The field studies reported here are the 1987 results of a larger five-year Canada−Newfoundland Mineral Development Agreement (CNMDA) project on the 'Metallogeny of the Central Mineral Belt, Labrador'. Preliminary field results from the 1987 study have been described elsewhere (Wilton and Brace, 1988).

The Central Mineral Belt (CMB) is a 260 by 75 km area, within central to coastal Labrador that is underlain by six Proterozic supracrustal sequences, *viz*., the Moran Lake, Lower Aillik, Upper Aillik, Bruce River, Letitia Lake and Seal Lake groups, ranging in age from *circa* 2000 Ma to *circa* 1300 Ma (Ryan, 1984). The Seal Lake Group, the youngest supracrustal sequence, hosts epigenetic copper occurrences ranging from native copper in slate and basalt, to chalcocite and bornite in slate, diabase sills and basalt flows.

The main objectives of the Seal Lake study were to, 1) document the numerous copper occurrences in the area *(this report)*, and 2) to sample the gabbro and diabase sills for petrogenetic modelling and definition of their Platinum-Group Element potential (Wilton, *this volume)*.

THE SEAL LAKE GROUP

Introduction

The Seal Lake Group (after Baragar, 1981, and Ryan, 1984) is a 1323 ± 92 Ma (Wanless and Loveridge, 1978) old unit that consists of six formations (see Figure 1) *viz*.; the Bessie Lake−Majoqua Lake Formation (quartzite or arkosic sandstone, and basalt); the Wuchusk Lake Formation (red to black shale, and minor quartzite and limestone); the Whiskey Lake and Salmon Lake formations (red shale, basalt and minor quartzite); the Adeline Island Formation (red to grey shale and quartzite with minor basalt), and the Upper Red Quartzite Formation. The middle sedimentary sequences (i.e., Wuchusk Lake, Whiskey Lake and Salmon Lake formations) contain numerous diabasic to gabbroic sills of the Naskaupi sills unit (Baragar, 1981). Throughout these six lithologic units, the group is host to over 250 Cu occurrences in quartz veins and/or shear zones (Brummer and Mann, 1961).

Previous Work

Most of the information available on the Seal Lake copper occurrences is in the form of unpublished mining company reports. Evans (1952) produced the first geological map of the Seal Lake Group, defined the stratigraphy of the group, and outlined the synclinal form of the unit. He also reported that the first discovery of native copper in the area was made in 1946 by geologists with Norancon Exploration Limited, south of Adeline Lake.

Robinson (1956) described briefly the geology of the Seal Lake Group and suggested that the group belonged within the Grenville geological province. Fahrig (1957) described the regional context and lithostratigraphy of the Seal Lake Group in greater detail and later (Fahrig, 1959) produced a 1:250,000 scale map of the regional geology.

Geologists from Frobisher Limited undertook extensive exploration and evaluation programs in the Seal Lake area during the years between 1950 and 1955 (Evans, 1951a,b; Robinson, 1953, 1954, 1955), which led to the discovery of numerous copper occurrences throughout the area. Their

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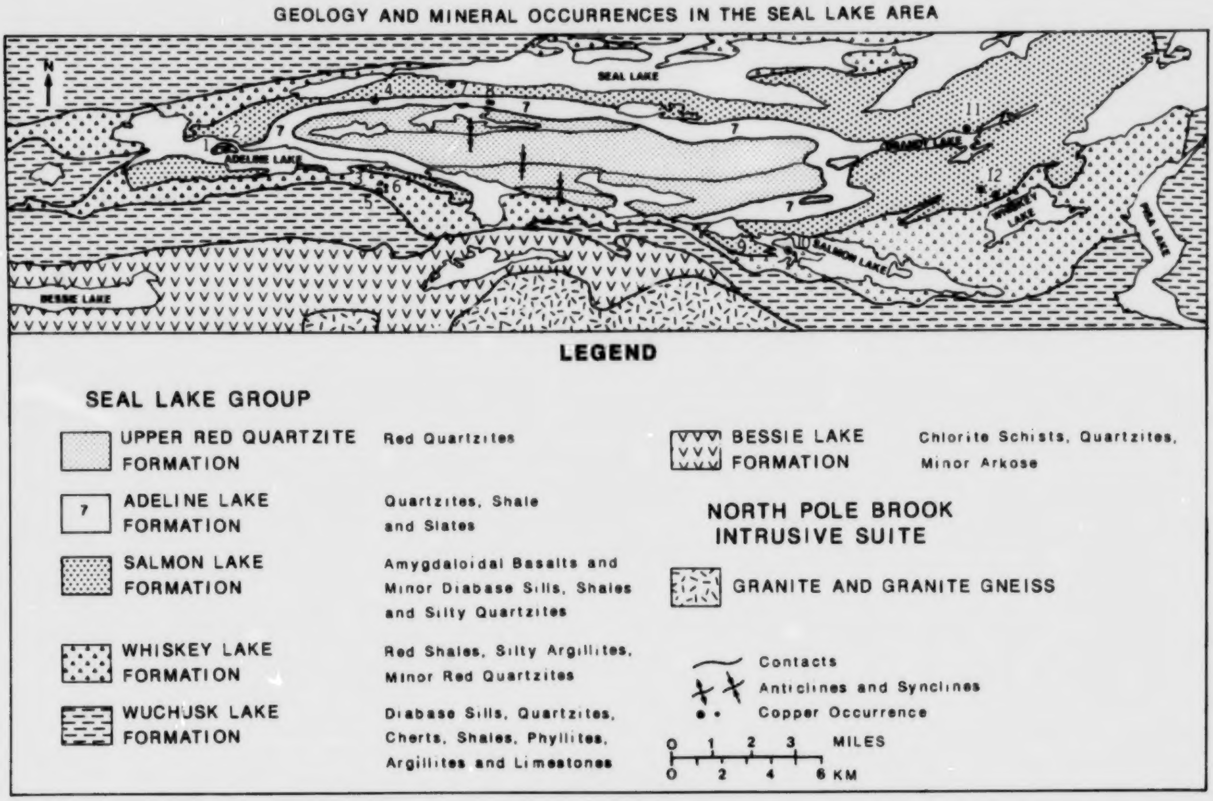
GEOLOGY AND MINERAL OCCURRENCES IN THE SEAL LAKE AREA
N
SEAL LAKE
ADELINE LAKE
BRANDY LAKE
WHISKEY LAKE
PIKE LAKE
SALMON LAKE
BESSIE LAKE
1
2
3
4
5
6
7
8
9
10
11
12
LEGEND
SEAL LAKE GROUP
UPPER RED QUARTZITE FORMATION
Red Quartzites
ADELINE LAKE FORMATION
Quartzites, Shale and Slates
SALMON LAKE FORMATION
Amygdaloidal Basalts and Minor Diabase Sills, Shales and Silty Quartzites
WHISKEY LAKE FORMATION
Red Shales, Silty Argillites, Minor Red Quartzites
WUCHUSK LAKE FORMATION
Diabase Sills, Quartzites, Cherts, Shales, Phyllites, Argillites and Limestones
BESSIE LAKE FORMATION
Chlorite Schists, Quartzites, Minor Arkose
NORTH POLE BROOK INTRUSIVE SUITE
GRANITE AND GRANITE GNEISS
Contacts
Anticlines and Synclines
Copper Occurrence
0 1 2 3 MILES
0 2 4 6 KM

Figure 1. *Geology of the Seal Lake Group and the Seal Lake Syncline (after Brummer and Mann, 1961). Copper occurrences are denoted by black circles; the larger black circles indicate showings that have been evaluated in detail by mineral exploration companies. The location of the various copper occurrences discussed in the report are shown as follows: 1) Adeline Island West Occurrence; 2) Adeline Island Showing; 3) NMI 13K/5 Cu018; 4) Ellis Showing; 5) South Adeline Mountain Showing; 6) NMI 13K/15 Cu187; 7) Seal Lake Main Showing; 8) Brian Prospects; 9) Showing No. 36; 10) Salmon Lake Main Showing; 11) Brandy Lake Showing; and 12) Whiskey Lake Showings.*

evaluations included data collected from trench-blasting, diamond-drilling, and electromagnetic and scintillometer surveys.

Kennco Explorations (Canada) Limited optioned the concession area from Frobisher Limited in 1956 and undertook a further extensive property evaluation program during the 1956 field season. The 1956 program involved detailed examination of the Seal Lake Main Showing, the Adeline Island Prospect, the North Adeline Lake Prospects, the South Adeline Mountain Prospect, and the North Whiskey Lake Prospects. These investigations included mapping, geophysical surveys and diamond drilling. A pioneering stream-sediment geochemical survey was also completed in this area as reported in Brummer (1960). As a result of the 1956 field work, Brummer (1957) recommended that no further work be carried out in the area as he felt the potential for ore-grade concentrations of copper did not exist in the syncline.

Brummer and Mann (1961) produced the definitive work to date on the stratigraphy, nomenclature and geological distributions of both lithologies and mineral occurrences within the Seal Lake Group. They also described four showings in detail (*viz.*, the Adeline Island, Ellis, and North Whiskey Lake prospects, and the Seal Lake Main Showing).

Baragar (1969) completed two chain and compass stratigraphic section lines across the Seal Lake Group. These lines provided a complete section through the stratigraphy of the Seal Lake Group on the northern limb of the syncline. Baragar (1974, 1981) used the analytical data to define the sills as transitional between alkaline and subalkaline tholeiites and the basalts as olivine normative plateau basalts.

Commencing in 1970, geologists from BRINEX Limited mounted an exploration program in the basin (Gandhi, 1970). The main impetus for the exploration was the postulated geological similarity of a sulphide-bearing grey-black slate unit, within the Adeline Island Formation, and the Nonesuch Shale Formation in the White Pine Copper Deposit of Michigan (Gandhi, 1972a). Evans (1952) had previously suggested that the Seal Lake copper deposits were similar to those of the Keweenawan Peninsula of Michigan. Brown (1971) completed a mineralographical study of samples from the Ellis and Adeline Island showings in order to ascertain more fully their similarities with the White Pine Deposit. He suggested that, although some features were very much alike, a definite correlation could not be made due to deformation.

The exploratory program undertaken by BRINEX Limited from 1970 to 1972 involved detailed geological mapping (to define the stratigraphy of the Adeline Island Formation throughout the Seal Lake Basin), geophysical surveys and diamond drilling (Gandhi, 1970, 1971, 1972a,b; Murthy and Gandhi, 1972). The final conclusion from the evaluation of the basin was that, though there were, 'several small pockets of disseminated copper–silver mineralization in grey shales, ... none of them is of size and grade to be of economic interest' (Gandhi, 1972a, p. 13). Exploration activity by BRINEX Limited personnel ceased following the 1972 field season and there has been no further mineral exploration in the area.

The most recent report on the copper occurrences was by Gandhi and Brown (1975), who described in detail five showings (*viz.*, the Ellis, Brian, Duck Lake South, Duck Lake North and Brandy Lake prospects; the main BRINEX Limited targets) that they defined as occurring solely within the shale and quartzite of the Adeline Island Formation. They modelled this formational unit as an analogue to the Nonesuch Shale and associated mineralization at White Pine, Michigan.

Description of copper occurrences examined in 1987. During the field program of 1987, over thirty-six copper occurrences were examined; ranging from malachite-stained quartz–carbonate veins to larger prospects that had been extensively drilled by the mineral exploration companies. The six prospects (copper occurrences numbered 2, 4, 7, 8, 11 and 12 on Figure 1) described by Brummer and Mann (1961) and Gandhi and Brown (1975) occur within the northern limb of the Seal Lake syncline in the Salmon Lake (the Seal Lake Main Showing) and Adeline Island formations. Thus, to date, there has been no single complete compilation of all the copper occurrences in the Seal Lake Group and in this report, the more significant of the other occurrences in the southern limb of the syncline (discovered in the early 1950's) will be described (occurrences numbered 3, 5, 6, 9 and 10 on Figure 1). The alpha-numeric designation, of copper occurrences described in the text, refers to that used on Map 83-27 (O'Driscoll and Noel, 1983).

Adeline Island West

This occurrence, at the western end of Adeline Island, is on the same stratigraphic level as the Adeline Island Showing. It is a small (20 by 2 m), trenched, copper showing, consisting of quartz veins up to 0.3-m thick, having associated chalcopyrite, pyrite, bornite and molybdenite. The veins occur in a grey quartzite unit, between the contact zone with a 2-m-thick black slate bed. Molybdenite was discovered at this site during the 1987 field season.

Adeline Island Showing

At this showing, bornite and chalcocite occur within quartz veins in the contact zone between a phyllitic grey-black siltstone and massive red to grey quartzite (Adeline Island Formation). The siltstone layer has an irregular distribution and thickness (ranging from $\leq$ 3 m to $<$ 5 cm) due to infolding within the quartzite. The siltstone itself has infolded red slate layers (3- to 5-cm thick), which are cross-cut by quartz veins. The whole sequence dips to the south, and red quartzite and shale immediately underlie the ore zone. Stratigraphically below the mineralization is red slate and basalt of the Salmon Lake Formation.

There were two periods of folding; an earlier isoclinal folding episode, which produced folds with vertical to subvertical fold axes, and the later episode, which produced the larger folds (isoclinal and overturned with axial planes dipping to the south) described by Brummer and Mann (1961). Axial-planar foliation associated with the D_1 isoclinal folding has, in places, completely transposed bedding into a parallel secondary fabric.

There are at least two generations of quartz veining at the showing. The copper sulphides are hosted by the first generation (folded and boudinaged) set, thus the mineralization was syn- or slightly post-D_1. The second generation, undeformed, quartz vein set often brecciates the host rocks.

Where quartzite layers have been isoclinally folded (as visible at Adeline Falls), the ductility contrast between quartzite and enclosing phyllite produced dilatant zones in the quartzite, which were subsequently filled with quartz (Plate 1). Quartz veins in similar zones contain bornite at the showings (Plate 2). In other parts of the showing, where the quartzite–phyllite contacts are more linear, the sulphide-bearing quartz veins are also linear, although sheared.

Parts of the quartzite, which host the most intense copper mineralization, often have a grey 'washed out' colour, suggesting that the copper-bearing hydrothermal fluids reacted with the host rocks removing Fe to form bornite.

This showing is currently being studied by P. Ivany as the basis of a CNMDA-supported B.Sc. (Hons.) dissertation at Memorial University of Newfoundland.

Seal Lake Showing No. 1– NMI [13K/5 Cu018]

This is Robinson's (1954) Showing No. 1, at the eastern end of Adeline Lake, which consists of a single, 28-m-long trench through south-dipping (75°/45S) metabasalt. Within the trench is a podiform quartz–carbonate–epidote vein(158°/63E), which contains hackly native copper, and

Plate 1. *Isoclinally folded quartzite (white) within green phyllitic schist (black). The quartzite here occurs in a fold nose. Note the quartz vein and phyllite-filled fractures cutting the more competant quartzite.*

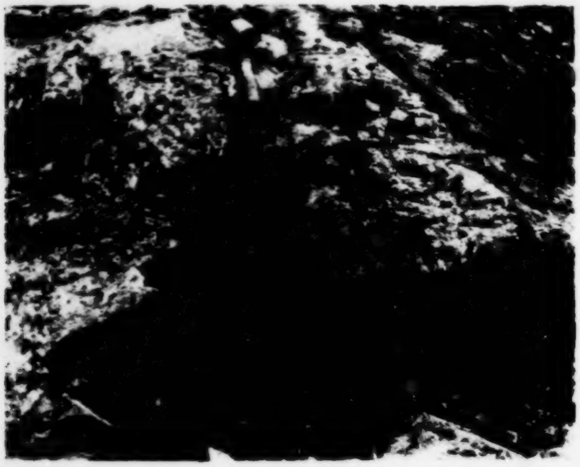

Plate 2. *Quartzite (massive grey rock in background) cut by black phyllite (parallel to the upper hammer handle) containing a bornite-bearing quartz vein at the Adeline Island Showing.*

associated malachite, azurite and cuprite. The vein appears to fill a fault–fracture system and is significant because it is the only example seen by this author in the Seal Lake area, wherein the vein system cuts the host rock at a high angle (i.e., in all other occurrences the vein systems are usually parallel or subparallel to the main S-fabric of the host rocks).

The Ellis Showing No. 26–NMI [13K/5 Cu002]

There are a number of different lithologies at this showing, and generally, mineralization seems to be contact-related. The main rock type is a grey quartzite (having visible, clear 1- to 2-m quartz grains interbedded with green siltstone, sandstone and minor fine-grained conglomerate. At one interval, the quartzite has an erosional contact (having a very thin basal conglomerate) indicating the units are upright. Stratigraphically above the quartzite (i.e., to the south), black slate and phyllite contain quartz–carbonate veins (up to 10-cm thick) having chalcopyrite, bornite and malachite (Plate 3). The quartzite contains bornite, chalcopyrite and/or pyrite stringers, which can account for up to 10 and 20 percent of the host rock (Plate 4).

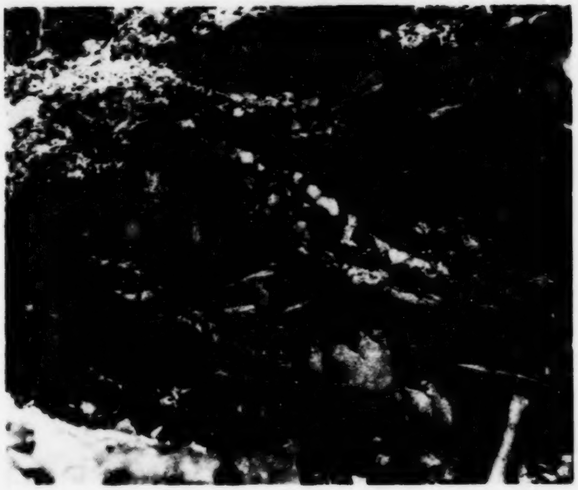

Plate 3. *Chalcopyrite (+ bornite) in quartz–carbonate vein (white in centre of photo) cutting green phyllitic schist at the stratigraphic top of the Ellis Showing.*

Stratigraphically below the quartzite, small limestone pods are interbedded with the green sandstone–siltstone, which ultimately pass downward into red slate.

This showing was reported (Robinson, 1954) to have grab sample assays of 2.86 percent Cu and 3.66 oz/t Ag from quartzite and 3.54 percent Cu and 2.8 oz/t Ag from slate. From one of the six blasted trenches, Robinson (1955) reported an assay of 2.5 percent Cu and 2.71 oz/t Ag over 4.5 m and an average grade of 1.35 Cu and 1.6 oz/t Ag across 4.2 m over a length of 60 m.

Based on the 1987 examinations, the mineralization is epigenetic with respect to the host rocks, the depositional site is a somewhat sheared sedimentary contact zone and the quartzite lenses are true sedimentary horizons. There is no dissemination of sulphide in the host rock, just stringers.

Plate 4. *Greenish-white quartzite (medium-grey massive rock) cut by stringers of bornite and chalcocite (black lines cutting quartzite) and quartz–carbonate veins (white) at the Ellis Showing.*

South Adeline Mountain Showing–NMI [13K/5 Cu019]

Robinson (1954) reported that his Showing No. 2 consisted of diabase underlain by hornfelsed argillite; shears along the contact contain quartz–carbonate veins, and pods and seams of bornite, chalcocite and chalcopyrite. According to Robinson (1955), the ore zone averages 1 percent Cu across a 5.1-m thickness over a 30-m length. Drilling by Kennco Limited (Brummer, 1957) revealed that the contact-related mineralization had a very erratic distribution.

This showing is exposed as two large blasted trenches and four channel-sample lengths along a north-facing cliff face (Figure 2). The cliff consists of a diabase sill overlying red sandstone and siltstone. The sill hornfelsed the Whiskey Lake Formation sedimentary rocks near the contact to the extent that they are dark green with flesh-coloured calc-silicate alteration patches. The mineralization consists of quartz–carbonate pods and disrupted veins, mainly within the metadiabase near the sedimentary contact, which contain impressive amounts of intergrown bornite and chalcocite. The widest pod is about 20 cm across (Plate 5).

South Adeline Lake No. 209A–NMI [13K/5 Cu187]

This showing occurs along the contact between the diabase cap of Adeline Mountain and the underlying hornfelsed shale of the Salmon Lake (or Whiskey Lake) Formation. Along the contact zone, the diabase is sheared and epidotized, often having malachite staining. The most intense malachite staining is associated with quartz–carbonate (± epidote) veins that contain chalcopyrite and pyrite (and minor bornite). Rare molybdenite was also found during this study, associated with copper minerals.

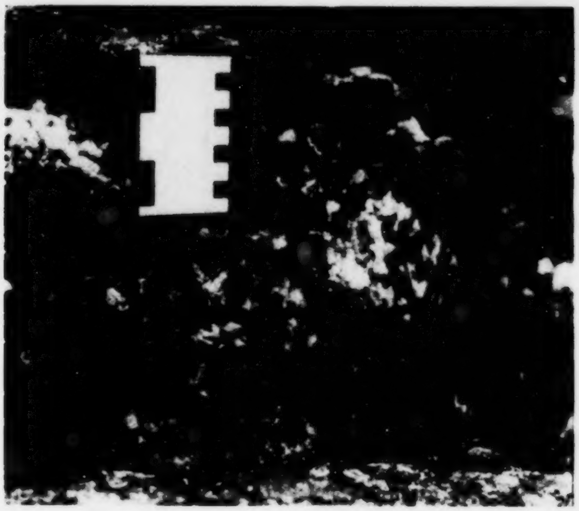

Plate 5. *Twenty-centimetre-wide quartz–carbonate vein (white) having abundant intergrown bornite-chalcocite (black in vein) from the South Adeline Mountain Showing; this particular vein occurs within the metadiabase.*

The Seal Lake Main Showing–NMI [13K/5 Cu001]

This showing consists of a 6- to 8-m-high cliff, 30 m from the south shore of Seal Lake, which is over 450-m long and has been trenched parallel to the cliff face for most of its length. Diabase occurs along the upper part of the cliff and is underlain by red slate, and in turn by the amygdaloidal basalt of the Salmon Lake Formation. At the contact, the diabase has hornfelsed the slate into a black slate. Within the red slate, quartz (+ carbonate) veins contain masses of native copper that form 'nuggets' up to 25-cm across. The veins trend parallel to the cliff face and are podiform within the cleavage planes of the slate. The slate wall rock is often brecciated and cemented by the veining. Quartz–epidote (± carbonate) veins within the diabase contain chalcocite and bornite, and rare native copper-bearing quartz–carbonate veins.

The showing resulted from a single mineralizing episode in which Cu-rich hydrothermal fluids migrated through the slate and diabase along fracture systems near their mutual contact. The oxidized nature of the slates (as indicated by their red colour) led to the precipitation of native copper. The more reduced nature of the diabase (magnetite-bearing and disseminated pyrite) led to precipitation of chalcocite and bornite.

The Brian Prospects–NMI [13K/5 Cu190 and 191]

There are two prospects, tentatively called the South and North zones (Gandhi and Brown's (1975) Brian Prospects 3a and 3b, respectively). The North Zone occurs in a 10-m-high cliff face, which has red slate at its base, grading upward into black and grey phyllite. The occurrence consists of

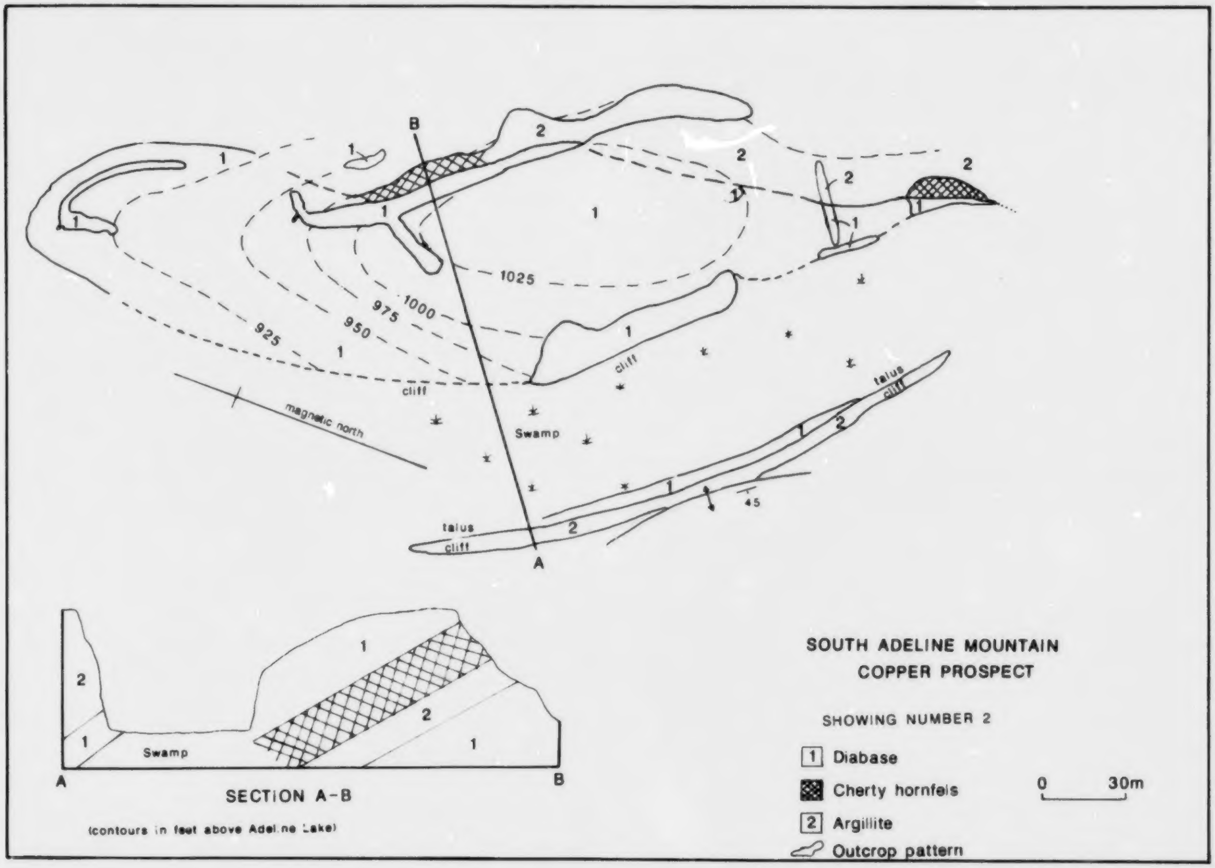

Figure 2. *Geological map of the South Adeline Mountain Showing (after Robinson, 1954). Podiform quartz–carbonate veins having bornite and chalcocite occur in the contact zone of the diabase and argillite; herein designated as cherty hornfels.*

(folded and/or boudinaged) quartz–carbonate veins within a 0.5-m-wide shear zone in green–grey phyllitic slate. The veins are up to 1.5-cm thick and contain chalcopyrite, bornite and pyrite.

At the South Zone, a 20-cm-wide quartz–carbonate vein occurs in grey–green quartzite, 0.6 m from a contact between the quartzite and black slate. The vein contains spectacular intergrowths of bornite and chalcopyrite. Pyrite stringers and cubes are present in both the slate and quartzite near the contact and there is minor malachite staining along the contact. Sedimentary facing directions are difficult to discern but it appears that the quartzite underlies the slate, and the mineralized zone is above the contact zone. In a more easterly trench (120 m from the main part of the zone) the quartzite is cut by quartz–carbonate veins containing bornite–chalcopyrite–pyrite (there is no slate in contact with the quartzite; i.e., the quartzite is the only rock in the trench).

West Salmon Lake No. 36–NMI [13K/5 Cu022]

This showing was discovered and described by Robinson (1954). It consists of a single 16-m-long stripped, and in places blasted, outcrop, along with smaller stripped outcrops extending for over 75 m west of the showing. This showing resembles the Adeline Island Showing in that it consists of quartz (+ carbonate) veins (up to 5-cm wide) along the lithological contact between a grey arkosic sandstone and chloritic schist. The veins contain bornite and malachite, and where veins are not present, intense malachite staining is present along the 75 m exposed contact zone. In the more western outcrops, veins contain pyrite. Robinson (1954) noted that channel samples across 0.75 m and 30 cm contained 0.83 percent Cu and 1.24 oz/t Ag, and 1.93 percent Cu and 0.22 oz/t Ag, respectively. During this study, minor amounts of molybdenite were found in the contact veins approximately 60 m west of the showing.

Salmon Lake Main Showing–NMI [13K/5 Cu017]

The showing occurs in an outcrop along the shore of Salmon Lake as a small blasted trench and has received relatively minimal evaluation. At this showing (Figure 3), epidotized and hematized basalt is cut by a native copper-bearing quartz–carbonate vein up to 0.3-m wide and over 20-m long. There is a smaller trench with native copper in

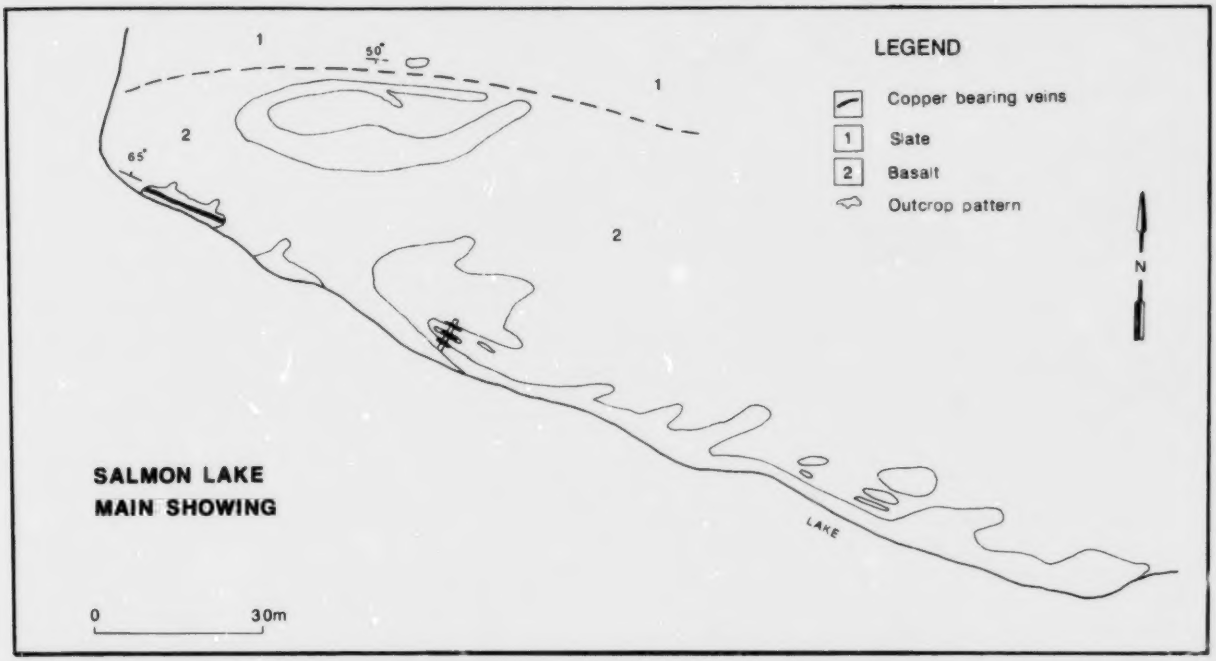

Figure 3. *Geology of the Salmon Lake Main Showing (after Robinson, 1953)*

quartz–carbonate veins, 40 m to the east. Robinson (1953) reported that grab sample from a 25-cm-thick section of the main vein contained 0.44 oz/t Ag and 16.72 percent Cu, and a chip sample across the eastern trench contained 0.07 oz/t Ag and 1.22 percent Cu.

There is another small occurrence 1 km to the west, which is seemingly an extension of the Salmon Lake Main vein system (Robinson, 1953). At this other occurrence, epidotized zones within chloritized and hematized basalt are cut by quartz–carbonate veins (up to 6-cm thick) that contain native copper and are malachite-stained. Robinson (1953) reported that two assays from this occurrence indicated 1.44 and 0.48 percent Cu across 0.8 and 1.2 m.

Brandy Lake Showings–NMI [13K/5 Cu029–032]

These showings are a series of small trenches spread over a 450 by 900 m area. The best mineralization occurs at the Brandy Main Showing and consists of a short trench and 3 by 5 m of blasted outcrop. There are a range of lithologies at this showing ranging from quartzite, to sandstone, and conglomeritic sandstone. Extensive chalcocite and bornite are associated with a sheared conglomeratic sandstone that has red (hematized) patches. Murthy and Gandhi (1972) describe the mineralization as a 0.6-m-thick, 1.5-m-long zone of chalcocite in grey slate at the contact with diabase and lava. They report grab sample assays of 7.5 and 10 percent Cu, and 0.17 oz/t and 0.26 oz/t Ag.

Approximately 195 m north-northeast of the main showing, red to pink quartzite contains small quartz veins associated with minor bornite. One hundred and thirty-five metres northeast of the main showing, diabase contains epidote veins having chalcopyrite, and quartz–carbonate (± epidote) veinlets containing bornite. Three hundred and ninety metres east of the main showing, quartz veins were found containing chalcocite and malachite in flesh-coloured phyllitic schist that are overlain by red slate, which in turn is overlain by quartzite.

Approximately 660 m northeast of the main showing, quartz–carbonate veins having chalcocite and malachite occur along the contact between diabase and underlying green phyllitic schist.

One hundred and twenty metres northwest of the main showing in an outcrop that contains the contact between two, somewhat schistose, basaltic lava flows (one hematized, the other chloritized with numerous amygdules), chalcocite and malachite containing specular hematite occur in carbonate–epidote patches and veins along the contact zone.

From the 1987 field survey, there is no evidence of copper sulphide minerals present in grey shales other than as epigenetic veins. Furthermore, the vein mineralization is found in all rock types and occurs in zones of structural weakness such as lithological contacts and hence it would appear that the copper mineralization in the Brandy Lake Showings was epigenetic.

North Whiskey Lake Showings–NMI [13K/5 Cu003]

There are two occurrences at this location, *viz.*, Mineral Occurrence (M.O.) 64 (after Brummer, 1957) to the south, and M.O. 150 to the north. Mineral Occurrence 64 consists

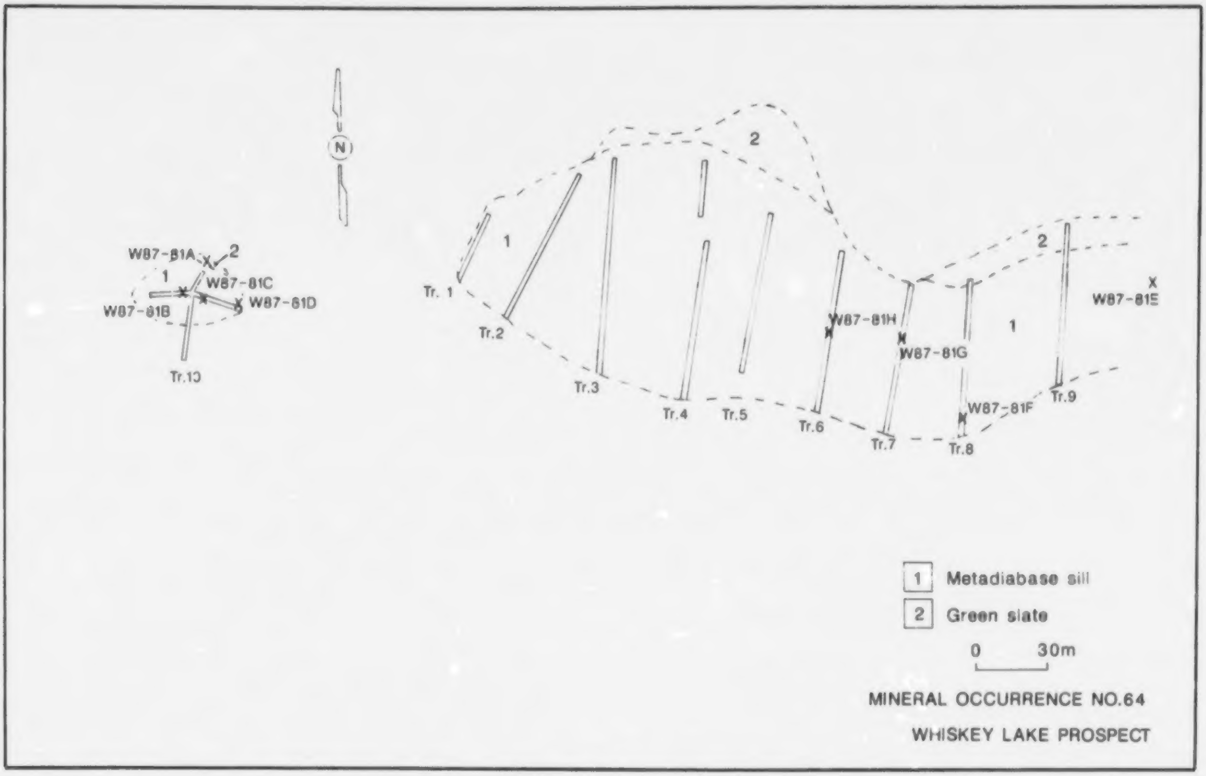

Figure 4. *Geology and sample location map of Mineral Occurrence No. 64, Whiskey Lake Showings (after Brummer, 1957).*

of eleven trenches across a diabase sill (Figure 4), which is underlain by grey slate–phyllite. There are a series of different vein systems cutting through the sill, including epidote veins and chalcopyrite and quartz–carbonate veins and bornite. The best mineralization occurs in trenches 10 and 11, where there are at least two generations of copper-bearing veins. The earlier set consists of epidote–quartz–carbonate veins containing chalcopyrite, and quartz veins having bornite, both of which are irregular and generally discontinous, and are associated with strong epidote alteration of the host diabase. The second set is a series of well-crystallized quartz–carbonate veins (up to 12-cm thick) containing chalcopyrite and chlorite.

On the northeastern edge of the outcrop grey slate–phyllite underlies the diabase (W87-81A, Figure 4). Close to the contact, the slate contains thin ($\leq$ 1-cm thick) carbonate veinlets having chalcopyrite–pyrite and malachite staining.

In the remaining trenches, the diabase contains scattered, erratic, and thinner epidote–quartz–carbonate veins having bornite, chalcopyrite and pyrite. Molybdenite occurs in a 15-cm-thick epidote vein in trench six (sample W87-81H, Figure 4).

Robinson (1955) called this occurrence Whiskey Lake No. 1 and described average assay values across the trenches as ranging from 0.08 percent Cu over 9 m to 1.6 percent Cu across 18 m.

Mineral Occurrence 150 is located approximately 300 m northwest of Mineral Occurrence 64. There are eight trenches at this showing. In some trenches, the rock looks like a chloritized sedimentary lithology (small quartz grains are visible, similar to those seen in quartzite at the Ellis Showing). This rock type is interbedded with purple-red slate. Mineralization consists of chalcocite disseminations and stringers within shear zones cutting the chloritized host rock. There are also minor quartz–carbonate veins having specular hematite, which seem unrelated to the copper mineralization. The best assay value from this occurrence yielded 7.11 percent Cu and 2.92 oz/t Ag over 1.5 m (Brummer, 1957).

Conclusions on the Mode of Origin of the Copper Occurrences

Host rocks to the copper occurrences in the Seal Lake Group include slate and phyllite of the Whiskey Lake Formation, red and green slate, shale, amygdaloidal basalts and diabase of the Salmon Lake Formation, the mixed sedimentary (slate, shale, phyllite and argillite) units of the Adeline Island Formation, black-grey shale of the Wuchusk Formation, and coarse to medium-grained diabase of the Naskaupi sills. In all cases, the copper mineralization is demonstrably epigenetic with respect to host rocks.

In general, native copper in quartz vein systems occur within red slate and hematized basalt units (although there are some examples in diabase). Bornite, and lesser chalcocite, occur in those vein systems within diabase, basalt or green to black phyllitic schist. Deposition was, therefore, seemingly related to the oxidation state of the host rock (or was predicated on the presence or absence of iron sulphide); in the oxidized (red) slates and basalts, copper precipitated from solution in its native (or oxidized form). In those units that contain primary sulphide, however, copper precipitated as a sulphide. The association of native copper with hematized basalt, further indicates the epigenetic nature of the mineralizing events as the hematization was obviously a result of post-depositional alteration.

Some units, such as slate in the Wuchusk Formation, phyllites in the Adeline Island Formation, and diabase in the Naskaupi sills, contain primary, syngenetic sulphides in the form of disseminated pyrite. Copper sulphide mineralization usually developed in proximity to these Fe sulphide-bearing units, thus, the fluids were essentially Cu-rich and oxidized. The fluids were focussed along zones of structural weakness, especially contact zones. As visible at the well exposed Adeline Island Showing, the mineralization appears to be syntectonic to slightly posttectonic with respect to D_1 isoclinal folding and shearing.

There are no substantial differences between copper mineralization in the Salmon Lake and Adeline Island formations, except the presence of native copper in the Salmon Lake rocks, which is apparently dependant on the oxidation state of the host rock (not on stratigraphic position). Both units host chalcocite, bornite, and chalcopyrite in diabase and sedimentary rocks.

Brummer and Mann (1961) suggested that copper mineralization was restricted to the upper portions of the Seal Lake Group because only the uppermost lava flows and sills contained sufficient copper to be concentrated epigenetically into shears and veins during later folding. Baragar (1981), though, reported lower copper contents within the stratigraphically higher Salmon Lake Formation when compared to lower volcanic units.

Gandhi and Brown (1975) examined the Adeline Island Formation and stated that the copper therein was derived from underlying volcanic rocks and transported to pyrite-bearing shales of the formation along 'red bed aquifers'. The copper-bearing fluids subsequently reacted with, and replaced, the pyrite, and precipitated copper sulphides in these shales (presumably syn-to slightly post-diagenetically). To support this model, these authors redefined the stratigraphy at the Adeline Island, Ellis and Whiskey Lake prospects (from that of Brummer and Mann (1961) and earlier workers) such that the host rocks to the mineralization were specified to be a single shale unit overlying a quartzite (the proposed aquifer) member of the Adeline Island Formation (i.e., stressing the singular ore bed comparison with the Nonesuch Shale).

Field observations during 1987 suggest that none of the red sedimentary units within the Whiskey Lake, Salmon Lake, or Adeline Island formations possessed sufficient porosities to have acted as 'red bed aquifers' (i.e., the units are not conglomeratic but impermeable shale or quartzite). Certainly, there is nothing distinctive about the middle quartzite member (Gandhi and Brown, 1975) of the Adeline Island Formation compared to other quartzite units in the upper parts of the Seal Lake Group. Furthermore, ore-fluid flow was demonstrably deformation-related, as mineralization only occurs in sheared rock along lithological contacts.

Determination of the actual source(s) of the copper and hydrothermal fluids awaits more detailed geochemical, isotopic and fluid inclusion studies (in progress). One interesting phenomenon that may reflect on the origin of the mineralizing components is the presence of molybdenite in at least four occurrences. Molybdenite would suggest magmatic involvement (i.e., a deeper seated granitic system) in the origin of the ore fluids. A similar model has also been suggested by Richards and Spooner (1986) for some of the native copper occurrences in the Keweenawan Peninsula, which they postulate (on the basis of fluid inclusion and carbon isotope evidence) to have been produced by the mixing of magmatic (high temperature and high salinity) fluids with cooler dilute meteoric waters.

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STUDIES IN THE SEAL LAKE GROUP, CENTRAL LABRADOR, PART II: GEOCHEMISTRY OF DIABASE SILLS

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ABSTRACT

The intermixed basaltic flow and continental sedimentary rocks of the Middle Proterozoic Seal Lake Group in central Labrador are intruded by the numerous diabasic to gabbroic sills of the Naskaupi sills unit. The major and trace-element geochemistries of thirty-five samples of these sills were analyzed to define, 1) the petrogenesis of the sills, 2) the geochemical signatures of copper-bearing ore fluids that flowed through the contact zones of some sills, and 3) the Platinum-Group Element (PGE) potential of the sills. The sills are transitional tholeiitic to alkalic in composition and resemble typical continental flood or plateau basalts. Those sills affected by the copper-bearing hydrothermal fluids have extreme depletions in K, Rb and Ba and slight enrichments in Sr. The sills do not appear to be a good prospect for PGE mineralization.

INTRODUCTION

As described in part I (Wilton, *this volume*), the Seal Lake Group consists of six mixed igneous and sedimentary formations (Brummer and Mann, 1961; Gandhi and Brown, 1975), the middle members of which (the Wuchusk Lake, Whiskey–Salmon Lake, and Adeline Island formations) are intruded by diabasic to gabbroic sills (Figure 1). These sills are so numerous in the Wuchusk Lake Formation, south of Adeline Lake that they constitute > 70 percent of the interval between the underlying Bessie Lake Formation and the overlying Whiskey–Salmon Lake Formation.

Baragar (1969, 1974, 1981) completed detailed petrographical and geochemical studies of the sills, north of Seal Lake on the northern limb of the Seal Lake syncline. In this present study, the sills south of Seal and Adeline lakes (i.e., in the southern limb of the syncline) were examined. The stratigraphy in the southern limb is compressed compared to that of the northern limb because of thrust faulting (Brummer and Mann, 1961), and the sills south of Seal Lake also contain numerous copper occurrences. Detailed prospecting (Brummer and Mann, 1961; Gandhi and Brown, 1975) has failed to reveal any significant occurrences north of the lake. Thus, the aim of this study was to compare the sills south of the lake and syncline, with those described by Baragar (1981) to the north.

PREVIOUS WORK

Evans (1952) described diabase sills as being intrusive into all formations of the Seal Lake Group. He also mentioned the very coarse grained nature of the sills near Adeline Lake.

Fahrig (1957) suggested that the diabasic to gabbroic sills are concentrated between the mafic volcanic rocks of the Bessie Lake Formation and the mixed red slate and basaltic volcanic members of the Salmon Lake Formation. He described the sills as being up to hundreds of feet thick, and associated with coarse plagioclase and pyroxene crystals (up to 2.5 cm in diameter), in the central parts of the sills. He also noted the presence of olivine during microscope studies.

Brummer and Mann (1961) subdivided the diabase sills into two different formations, *viz.*, those within the Wuchusk Lake Formation and those within the Salmon Lake Formation. The Wuchusk sills, along the southern limb of the syncline, are metadiabase due to saussuritization of plagioclase, serpentization of the olivine, and chloritization and epidotization of augite. The Salmon Lake sills are also described (*ibid.*) as metadiabase, but more closely resemble the basalt flows within their host unit, than the diabase sills in the Wuchusk Lake Formation.

Roy and Fahrig (1973) suggested a five-fold subdivision of the Seal Lake Group in which Unit 2, a volcanic and sandstone unit (the Majoqua Formation), and Unit 4, a volcanic and shale unit (the Salmon Lake Formation), were seperated by Unit 3, a diabase with sandstone and shale unit (i.e., the Wuchusk Lake Formation). To these authors, diabase intrusion between the two periods of volcanism represented the continuation of the igneous activity. They supported this supposition with geochemical data, which 'suggested a common magma source' (Roy and Fahrig, 1973, p. 1291) and they defined the volcanics and diabase as undersaturated tholeiites.

Gandhi and Brown (1975) stated that the diabase sills intrusive into the Adeline Island, Salmon Lake and Wuschusk formations were related to Salmon Lake Formation magmatic activity. They suggested that the diabase unit, intrusive into the Adeline Island Formation sedimentary rocks north of Brandy Lake, was actually posttectonic (post-Grenville), and

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This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

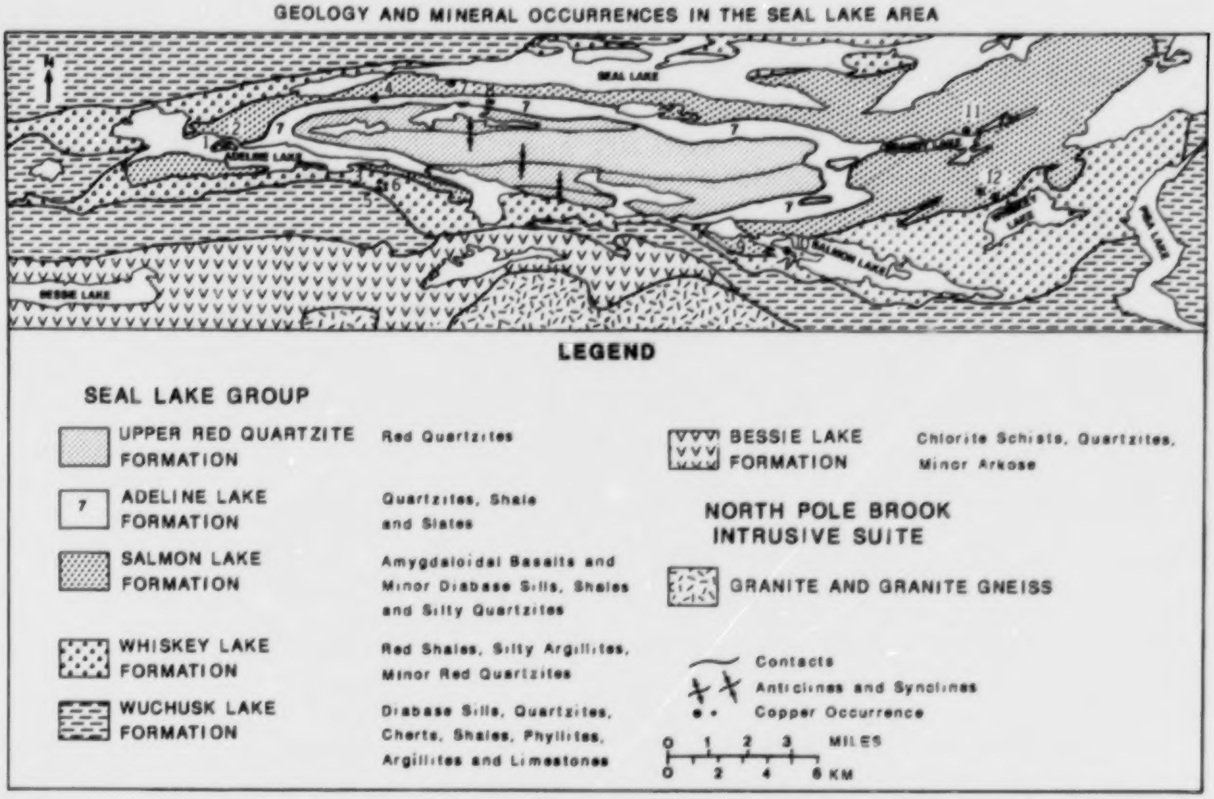
GEOLOGY AND MINERAL OCCURRENCES IN THE SEAL LAKE AREA
N
SEAL LAKE
ADELINE LAKE
BESSIE LAKE
SALMON LAKE
LEGEND
SEAL LAKE GROUP
UPPER RED QUARTZITE FORMATION
Red Quartzites
7
ADELINE LAKE FORMATION
Quartzites, Shale and Slates
SALMON LAKE FORMATION
Amygdaloidal Basalts and Minor Diabase Sills, Shales and Silty Quartzites
WHISKEY LAKE FORMATION
Red Shales, Silty Argillites, Minor Red Quartzites
WUCHUSK LAKE FORMATION
Diabase Sills, Quartzites, Cherts, Shales, Phyllites, Argillites and Limestones
BESSIE LAKE FORMATION
Chlorite Schists, Quartzites, Minor Arkose
NORTH POLE BROOK INTRUSIVE SUITE
GRANITE AND GRANITE GNEISS
Contacts
Anticlines and Synclines
Copper Occurrence
0 1 2 3 MILES
0 2 4 6 KM

Figure 1. *Geology of the Seal Lake Group (after Brummer and Mann, 1961). Key to deposit numbers and copper occurrences have been described and discussed. Most of the area underlain by the Wuchusk Lake Formation consists of diabasic Naskaupi sills (> 70 percent south of Adeline Lake). There are also numerous diabase sills in the Salmon Lake Formation and a few in the Adeline Island and Whiskey Lake formations.*

therefore unrelated to the 'deformed' diabase sills stratigraphically lower inthe Seal Lake Group.

Baragar (1981), mainly describing the Seal Lake Group north of Seal Lake, grouped all the sills into a new unit that he called the Naskaupi sills. He termed the sills, dolerite, and described them as intrusive into the Wuchusk Lake, Salmon Lake and Adeline Island formations. Baragar (1981) suggested that the sills intruded the thick accumulation of sedimentary rocks in the Seal Lake basin just prior to renewed volcanism, represented by the Salmon Lake Formation. He further defined the sills as being somewhat transitional between alkaline and subalkaline tholeiitic in composition, and that they could be termed the products of potassic or K-rich basalt magmas.

Meyers and Emslie (1977) and Emslie (1980) directly correlated the Naskaupi sills with the regionally extensive Harp dykes. These latter dykes were defined by these authors as having a chemical composition transitional between tholeiitic and alkalic. Emslie *et al.* (1984) similarly correlated the Mealy dyke swarms of the Grenville Province with the Seal Lake (Naskaupi sill) diabases. Wiebe (1985) suggested that dykes cutting the Nain Anorthosite Complex were compositionally and temporally similar to the Harp dykes and, hence, the Naskaupi sills.

Description of the sills

The diabase sills are best exposed in two sets of steep cliffs that trend east–west through the area as a ridge along the southern shoreline of Seal Lake, and as a prominant ridge extending from the southern shore of Adeline Lake to the southern shore of Salmon Lake (Figure 1); the former ridge rises up to 200 m above Seal Lake and the latter up to 300 m above Adeline Lake (at Adeline Mountain). In both cases, the ridges are capped by diabase sills, and there are thinner sills present in the Seal Lake cliff face below the cap. Adeline Mountain (or Wuchoo Mountain, Brummer and Mann, 1961), at > 540 m above sea level is the highest hill in the area, and is a perfect example of the typical diabase profile (Plate 1), wherein Whiskey Lake Group slates are 'capped' by 120 m of diabase dipping gently to the north.

Exposed contacts are often difficult to find due to vegetation cover, but locally, very fine grained phases are

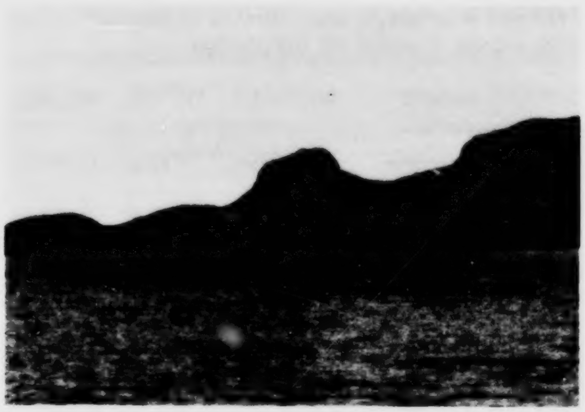

Plate 1. *View to the east along Adeline Lake with Adeline (Wuchoo) Mountain in the centre. This monadnock is capped by a diabase sill (shown by the change in slope) and underlain by hornfelsed shale; the ridge along the right is also capped by diabase.*

found within coarse layers implying the intrusion of sills within sills. Sedimentary lithologies (especially slates) are extensively hornfelsed in contact regions (in places, red slate is hornfelsed into black slate). Well developed columar jointing is evident in some of the thicker sills (Plate 2).

Plate 2. *Columnar jointing in diabase sill intrusive into the Salmon Lake Formation south of Seal Lake.*

The coarsest grained diabase, which occurs south of Adeline Lake, contains pyroxene crystals up to 2 by 2 cm, plagioclase laths up to 0.5 by 2.5 cm, together with magnetite blades (up to 2 cm by 3 mm) and rare, disseminated pyrite. Diabase, which is intrusive into the Salmon and Adeline Lake formations, is finer grained, more equigranular, and displays visible epidote alteration. As described in Wilton (*this volume*), the contact zones of the sills are the loci for quartz–carbonate veins containing copper sulphide mineralization.

Geochemistry

Geochemical analyses have been completed on thirty-five diabase–gabbro samples of the Seal Lake Group (Table 1) from the southern limb of the syncline (south of Seal Lake). Major-element analyses were done by atomic absorption techniques, and the trace-element determinations by XRF analysis of pressed powder pellets (Longerich and Veinott, 1986) at Memorial University.

One of the reasons for sampling the diabase sills was to define their PGE potential. Analysis for these elements has yet to be performed. However, other geochemical data from the Naskaupi sills tend to downplay their potential for significant primary (or orthomagmatic) concentrations of these elements. In all known examples of PGE concentrations within undeformed mafic to ultramafic rocks (the Naskaupi sills sampled are undeformed), platinoid minerals are associated with either chromite-bearing horizons and/or nickeliferous sulphides (e.g., Dillon-Leitch *et al.*, 1986; Irvine *et al.*, 1983; Macdonald, 1987). However, Ni and Cr concentrations within the Naskaupi sills are quite low, with Cr contents generally < 200 ppm (one sample had 317 ppm) and Ni also mostly < 200 ppm, implying that Cr and Ni-bearing phases are substantially lacking. The spinel present within the sills is obviously magnetite to titanomagnetite (as shown by the elevated TiO_2 contents of up to 3.8 wt percent).

In terms of some classic variation diagrams, the Naskaupi sills have a transitional subalkaline to alkaline (Figure 2) tholeiitic (Figure 3) composition. On Pearce and Cann's (1973) Ti *versus* Zr discrimination diagram (Figure 4a), the sills plot partly in the ocean-floor basalt (OFB) field, but are also so enriched in Zr and Ti that half of the samples plot outside the defining fields. However, on the Zr–Ti–Y discrimination diagram (Figure 4b) from Pearce and Cann (1973), the sills plot as transitional between calc-alkali and within plate basalts.

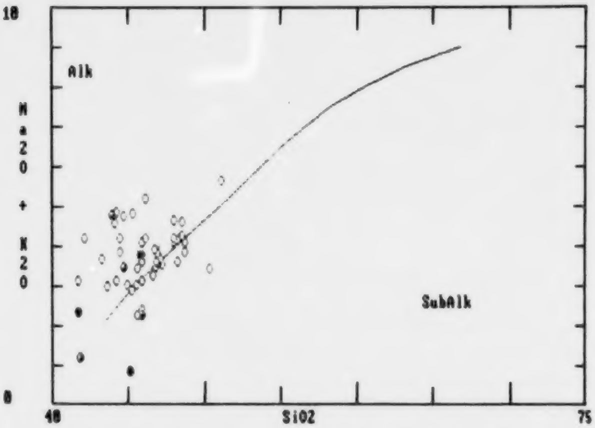

Figure 2. *Total alkalies vs. silica diagram for Naskaupi sill samples. Fields for alkaline (Alk) and subalkaline (SubAlk) magmas are from Irvine and Baragar (1971). Open circles are normal diabase, half-filled circles are of diabase having visible copper mineralization, filled circles are of diabase from near copper occurrences but are not actually copper or vein-bearing, the * symbol represents Baragar's (1981) average Nasakupi sill composition.*

Table 1. Geochemical data for Naskaupi sills; major oxide data reported in weight %; trace elements in ppm; n.d. = not detected; – = not analysed; Bar-avdol = Baragar's (1981) average Naskaupi sill composition.

	W87-105	W87-16	W87-46B	W87-71A	W87-31A	W87-93D	W87-46A	W87-81I	W87-81D
SiO_2	47.20	35.70	45.10	45.90	46.10	46.90	41.80	43.90	45.90
TiO_2	1.88	2.24	1.80	1	1.52	3.60	2.04	1.68	1.24
Al_2O_3	14.60	18.50	16	18.30	15.20	11.80	17.20	17.40	16.50
FeO	13.28	16.50	13.15	10.55	12.58	17.66	13.60	13.64	12.12
MnO	0.19	0.19	0.19	0.14	0.22	0.24	0.20	0.18	0.17
MgO	5.28	6.69	6.28	10.06	6.64	4.71	6.82	7.51	6.66
CaO	8.82	11.08	9.28	9.68	6.96	9.16	9.28	4.24	9.22
Na_2O	2.63	1.29	0.84	1.95	2.16	2.68	1.18	3.93	2.24
K_2O	0.90	n.d.	n.d.	0.43	3.04	0.84	n.d.	0.85	0.02
P_2O_5	0.25	0.22	0.22	0.11	0.21	0.24	0.22	0.24	0.20
Total	98.03	92.78	92.78	98.12	94.63	97.83	92.34	93.57	94.27
H_2O	-	-	-	-	-	-	-	-	-
CO_2	-	-	-	-	-	-	-	-	-
LOI	2.35	4.66	4.09	0.72	2.77	0.76	4.57	4.17	3.50
Mg #	41	42	46	63	48	32	47	50	49
Cr	65	148	108	144	101	n.d.	122	95	73
Ni	38	98	124	224	67	n.d.	138	122	156
Co	-	-	-	-	-	-	-	-	-
Sc	-	-	-	-	-	-	-	-	-
V	256	336	267	106	237	588	289	197	203
Cu	102	100	177	40	68	219	8345	71	72
Pb	n.d.	26	n.d.	n.d.	n.d.	2	n.d.	1	n.d.
Zn	103	122	108	54	103	128	127	128	102
Sn	-	-	-	-	-	-	-	-	-
K	7471	n.d.	n.d.	3569	25235	6973	n.d.	7056	166
Rb	17	n.d.	n.d.	9	81	17	n.d.	20	n.d.
Ba	364	1	66	106	517	354	n.d.	311	n.d.
Sr	264	723	023	273	193	215	1036	110	558
Ga	18	25	20	17	22	23	18	18	16
Nb	11	10	8	6	7	11	10	8	8
Zr	168	175	157	65	128	157	173	130	133
Ti	11271	13429	10791	5995	9112	21582	12230	10072	7434
Y	35	34.8	27	10	33	41	29	28	25
Th	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
U	3	5	n.d.	n.d.	3	3	n.d.	5	6
La	14	6	13	1	8	8	7	9	4

According to Condie *et al.'s* (1987) TiO_2 *versus* Zr discrimination diagram (Figure 5), the Naskaupi sill samples all plot in the combined midocean ridge–continental rift–oceanic-island basalt field similar to their Superior Province dyke data. The Naskaupi sills more closely resemble Condie *et al.'s* (1987) Archean and Early Proterozoic dykes than their Middle Proterozoic dykes, which are more enriched in incompatible elements. This later (Middle Proterozoic) enrichment was modelled (*ibid.*) as the result of assimilation of more evolved continental crust.

All the diabase samples collected for this study have similar compositions and generally exhibit coherent groupings on trace-element variation diagrams (e.g., Figures 2 to 5). The compositions are the same as those defined by Baragar (1981). Samples of diabase associated with epigenetic copper occurrences, however, have distinctive variations from the main group, for example Sr is increased (Figure 6) and SiO_2 decreased (Figure 7).

The [Mg] (where [Mg] = $(Mg/(Mg+Fe^{+2})$ and $Fe^{+3}/Fe^{+2} = 0.15$) values for the samples range from 0.19 to 0.62, but are generally less than 0.5 (Figure 7). Such low [Mg] values (< 0.68) coupled with the low Ni and Cr contents are typical of evolved lavas that have undergone extensive fractionation prior to intrusion (e.g., Goldberg *et al.*, 1986; Wiebe, 1985).

Table 1. *(Continued)*

W87-106	W87-74	W87-117	W87-88C	W87-71B	W87-80	W87-88A	W87-78A	W87-78B	W87-77B
48.20	45.90	48.70	46.90	46.80	45.90	48	45.90	48	51.10
1.72	1.20	2.32	3.80	1.72	3.04	2	3.80	2.92	2.80
14.70	17.50	13.50	11.20	15.90	10.60	12.90	10.60	11.20	12.60
11.10	11.60	14.28	18.45	12.65	14.88	16.57	18.74	16.77	15.21
0.17	0.16	0.20	0.26	0.21	0.22	0.23	0.28	0.26	0.26
6.17	8.41	5.19	4.12	6.82	6.30	3.71	3.54	3.80	2.02
9.54	9.04	7.44	8.58	9.58	11.16	8.08	8.08	8.80	7.06
2.80	2.54	2.99	2.61	3.02	2.39	2.74	2.75	3.64	4.24
1.36	0.58	0.86	1.18	0.45	1.20	1.44	1.34	0.98	1.40
0.02	0.17	0.27	0.35	0.22	0.30	0.33	0.68	0.63	0.88
95.78	97.10	95.75	97.45	97.37	95.99	96	95.71	97	97.57
-	-	-	-	-	-	-	-	-	-
-	-	-	-	-	-	-	-	-	-
2.07	0.10	2.88	0.55	1.95	0.90	1.13	1.92	1.29	1.12
50	56	39	28	49	43	29	25	29	19
76	47	53	n.d.	216	67	70	n.d.	29	19
69	132	25	n.d.	61	16	n.d.	n.d.	n.d.	n.d.
-	-	-	-	-	-	-	-	-	-
-	-	-	-	-	-	-	-	-	-
234	136	259	511	213	45	194	154	259	65
72	53	104	216	n.d.	253	32	170	178	55
n.d.	n.d.	2	n.d.	n.d.	1	4	n.d.	n.d.	1
81	68	112	143	75	93	133	194	164	169
-	-	-	-	-	-	-	-	-	-
11289	4815	7139	9795	3735	9961	11953	11123	8135	11621
27	13	21	33	8	26	43	47	31	36
677	266	350	443	224	333	473	521	389	494
244	311	204	220	285	186	239	228	302	243
19	20	20	25	19	19	21	24	22	27
8	5	11	13	9	11	12	20	18	28
129	86.8	212	197	111	153	194	259	259	418
10311	7194	13908	22781	10311	18225	11990	22781	17505	16786
30	18	42	47	22	37	45	56	58	78
n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
n.d.	n.d.	4	n.d.	4	3	6	2	5	1
3	5	16	16	5	11	27	24	24	43

Spidergrams

A rapid method for evaluating overall geochemical distributions and heterogeneity of a group of rocks is acheived with construction of extended trace-element variation diagrams (or 'spidergrams' after Thompson *et al.*, 1984). Figure 8 (a to h) consists of a number of these diagrams for all the Naskaupi sill samples analyzed (they are ratioed to Primitive Mantle values of Taylor and McLennan (1981) and Jenner *et al.* (1987)). There is a significant chemical coherence between all the samples from this study, and Baragar's (1981) average dolerite plots within this grouping. The high K (and Rb) nature of these sills as defined by Baragar (1981) is also borne out.

The overall patterns for the Naskaupi sills are, not surprisingly, quite similar to those for continental flood basalts (Figure 8g and h) from Thompson *et al.* (1983, 1984). Figure 8g shows typical patterns for continental flood basalts that resemble ocean-island basalts; Thompson *et al. (op. cit.)* suggest these magmas represent uncontaminated melts from ocean-island basalt mantle. Figure 8h illustrates Thompson *et al. (op. cit.)* patterns for continental flood basalts that have been variably contaminated by assimilation of continental crust upon magma ascent through the crust.

In general, the Naskaupi sills are intermediate between these two types of continental flood basalts. Compared to the ocean-island basalts and the continental flood basalts, the

Table 1. *(Continued)*

	W87-81E	W87-63	W87-72B	W87-42	W87-88B	W87-36	W87-16A	W87-77A	W87-31B
SiO_2	44.70	46.60	44.20	43.60	48.20	74.70	41.70	46.90	45.90
TiO_2	1.44	1.72	1.84	1.36	2.44	0.12	2.24	2.32	1.44
Al_2O^*	16.50	15.60	16.90	16.20	12.10	13.60	16.20	15.60	15.30
FeO	12.65	12.27	11.90	12.10	15.02	0.06	15.15	11.01	12.65
MnO	0.20	0.17	0.16	0.16	0.22	n.d.	0.21	0.19	0.17
MgO	6.89	6.35	7.98	10.08	5.04	0.02	5.75	5.57	7.02
CaO	8.56	9.32	9.54	8.28	9.54	0.18	9.28	9.68	7.86
Na_2O	2.90	2.51	2.44	2.42	2.58	3.22	2.19	2.89	1.79
K_2O	0.56	0.75	0.68	0.54	1.02	6.98	0.14	1.03	1.98
P_2O_5	0.19	0.21	0.24	0.16	0.29	0.02	0.27	0.40	0.16
Total	94.59	95.50	95.88	94.90	96.45	98.90	93.13	95.59	94.27
H_2O	-	-	-	-	-	-	-	-	-
CO_2	-	-	-	-	-	-	-	-	-
LOI	3.19	2.73	1.26	3.33	0.99	0.10	3.95	1.30	2.88
Mg #	49	48	54	60	37	37	40	47	50
Cr	66	76	80	53	196	n.d.	94	120	77
Ni	124	106	159	214	4	n.d.	41	50	93
Co	-	-	-	-	-	-	-	-	-
Sc	-	-	-	-	-	-	-	-	-
V	231	234	192	164	417	239	293	270	233
Cu	73	84	70	4	133	n.d.	118	97	72
Pb	5	n.d.	1	n.d.	n.d.	9	18	n.d.	1
Zn	101	94	73	76	119	139	142	84	97
Sn	-	-	-	-	-	-	-	-	-
K	4649	6226	5645	4483	8467	57934	1162	8550	16435
Rb	14	17	14	20	28	75	3	25	51
Ba	183	282	215	89	363	549	393	312	469
Sr	330	306	295	234	231	638	508	271	205
Ga	17	22	19	16	23	26	21	20	18
Nb	8	7	8	7	10	23	11	13	7
Zr	133	125	115	88	166	465	194	162	114
Ti	8633	10311	11031	8153	14628	719	13429	13908	8633
Y	26	26	23	22	39	91	36	36	26
Th	n.d.	n.d.	n.d.	n.d.	n.d.	6 0	n.d.	n.d.	n.d.
U	n.d.	n.d.	6	n.d.	n.d.	8	n.d.	n.d.	4
La	5	11	4	n.d.	12	48	16	11	6

Naskaupi sills have variably lower contents of K, La, Nb and Sr (the Sr differences may reflect different degrees of plagioclase fractionation). The La/Nb ratios of the Naskaupi sills lie between 0.5 and 2, which according to Thompson *et al.* (1983, 1984) is the general field of continental flood basalts. Continental flood basalts that resemble ocean-island basalts would normally have La/Nb ratios of about 1. According to Thompson *et al.* (1983, 1984) addition of normal crustal material to an ocean-island basalt—continental flood basalt melt, should provide a relative increase of K and La and decrease of Nb in the magma. Therefore, it would appear that if the parental magmas of the Naskaupi sills were contaminated, then the contaminant material would have to have been depleted (e.g., granulite) crust (i.e., something from which these typical mobile (fusible) components had been already removed).

Such a conclusion is supported by the isotopic data of Ashwal *et al.* (1987), which demonstrated only minor upper sialic crust contamination of the Harp dykes parental magma (i.e., correlatives of the Naskaupi sills). According to Ashwal *et al.* (1987) the actual contaminants to the mantle-derived parental magma would have been depleted in Rb, U—Pb and Th—U, and as such, probably represent Archean gneiss (by inference, this crust was probably granulitized—hence assimilation by the mafic magma would provide minimal addition of the typical upper crustal elements).

Table 1. *(Concluded)*

W87-72C	W87-65	W87-60C	W87-71E	W87-92	W87-71C/D	W87-71F	Bar-avdol
45.60	48	45.60	48.50	48.70	47.10	48.50	45.77
1.64	2.24	1.68	2.28	2.32	1.80	2.72	2.03
18.20	15.20	15.10	14.40	12.90	16	14.10	17.05
10.73	14.01	12.74	11.82	14.32	11.05	12.59	12.26
0.16	0.20	0.18	0.18	0.22	0.18	0.18	0.18
6.91	4.40	6.68	4.62	4.77	6.33	4.81	7.08
9.84	8.20	9.76	10.48	9.50	10.40	10.40	9.02
2.69	2.84	1.91	3.48	3.11	2.92	3.28	2.96
0.75	1.24	0.33	1.11	0.99	0.75	0.96	0.82
0.24	0.25	0.22	0.51	0.32	0.21	0.28	0.25
96.76	96.58	94.20	97.38	97.15	96.24	97.82	97.42
-	-	-	-	-	-	-	3.26
-	-	-	-	-	-	-	0.13
1.29	2.47	2.98	1.08	1.16	1.29	1.04	-
53	36	48	41	37	51	41	51
69	150	79	244	311	190	119	109
124	17	120	4	5	52	9	117
-	-	-	-	-	-	-	53
-	-	-	-	-	-	-	28
164	264	228	287	350	223	415	287
61	95	75	80	170	80	127	156
4	n.d.	1	n.d.	2	n.d.	6	5
69	111	100	69	128	62	75	185
-	-	-	-	-	-	-	1.61
6226	10293	2739	9214	8218	6226	7969	6807
13	26	6	23	25	15	21	n.d.
233	321	126	3471	452	221	287	223
300	271	338	263	272	270	240	271
20	21	18	21	23	18	21	32
9	10	7	13	10	9	12	-
122	162	132	170	170	113	174	155
9832	13429	10072	13669	13908	10791	16306	12170
23	35	24	38	37	25	36	31
n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	-
5	n.d.	1	n.d.	1	n.d.	2	-
10	9	13	22	16	60	9	-

As a final note, samples W87-60C and W87-63 were collected from the diabase ridge north of Brandy Lake. As shown on Figure 8b and e, these samples are exactly the same geochemically as the other sill samples. Thus, Gandhi and Brown's (1975) supposition that the sill north of Brandy Lake is post-Grenville, and therefore unlike the other sills, is probably incorrect.

Figure 9a and b are spidergrams of diabase samples that were mineralized with copper sulphide, or are from the vicinity of mineralization (the values for Baragar's (1981) average dolerite are also plotted on this diagram for comparison). This diagram illustrates the great changes in geochemistry brought about by reaction of the ore-forming fluids with the host diabase.

The unusual geochemical composition (and hence pattern) for sample W87-36 (Figure 8) is not as readily apparent. This sample was collected from a small island in Salmon Lake, wherein a diabase unit contained abundant disseminated pyrite and some minor quartz veining. It may actually reflect an earlier mineralizing event more closely related to magmatism (or a later event). Its main features are increased La and Zr and decreased Ti contents.

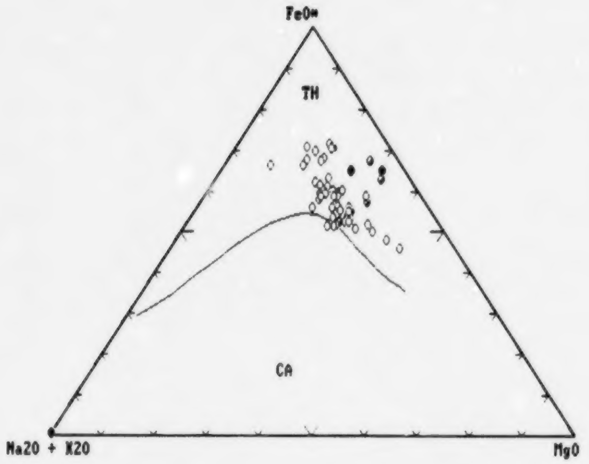

Figure 3. *AFM diagram for Naskaupi sill sample fields for tholeiitic (TH) and calc-alkaline (CA) magmas from Irvine and Baragar (1971) (the symbols are described in Figure 2).*

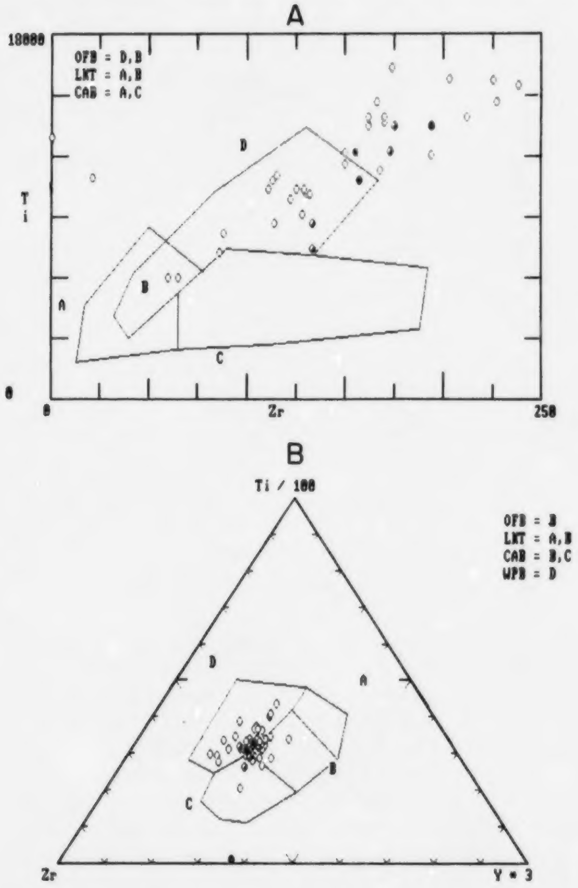

Figure 4a and b: *Trace-element discrimination diagrams (from Pearce and Cann, 1973) for Naskaupi sill samples. OFB = ocean floor basalts; LKT = low K tholeiites; CAB = calcalkaline basalts; and WPB = within plate basalts (symbols as in Figure 2).*

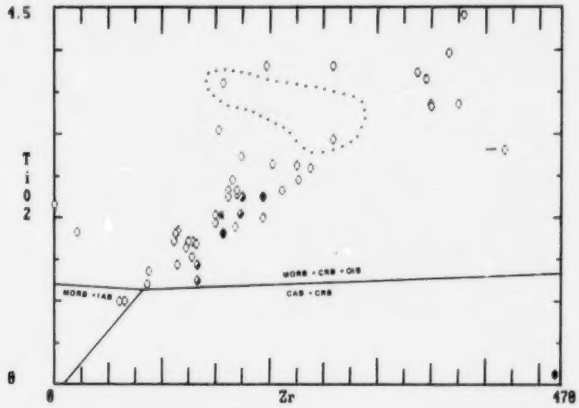

Figure 5. *TiO_2 vs. Zr plot for Naskaupi sill samples (symbols as in Figure 2). Subdivisions from Condie et al. (1987). MORB = mid-ocean ridge basalt, IAB = island-arc basalt, CRB = continental rift basalt, and OIB = ocean-island basalt. The dotted area represents Condie et al.'s field of Middle Proterozoic diabase dykes.*

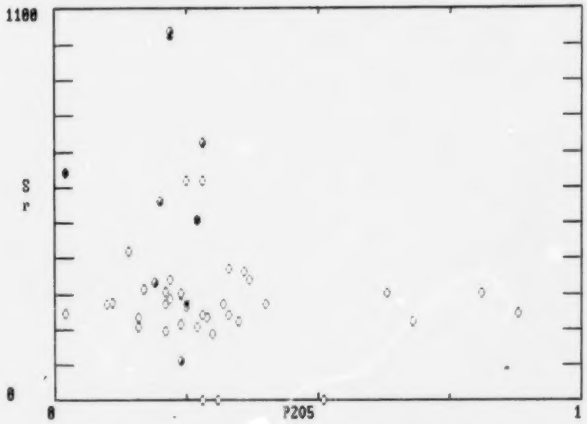

Figure 6. *Sr vs. P_2O_5 variation diagram for Naskaupi sills; the sills associated with copper mineralization have enriched Sr contents.*

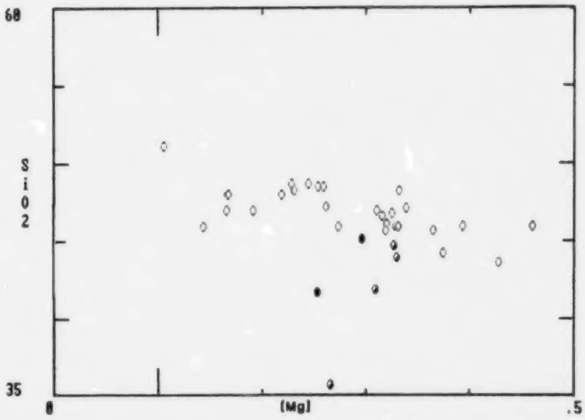

Figure 7. *SiO_2 vs. [Mg] for the Naskaupi sills; all samples have [Mg] of less than 0.6 (i.e., have undergone significant fractionation) and diabase samples associated with copper mineralization have decreased SiO_2 contents.*

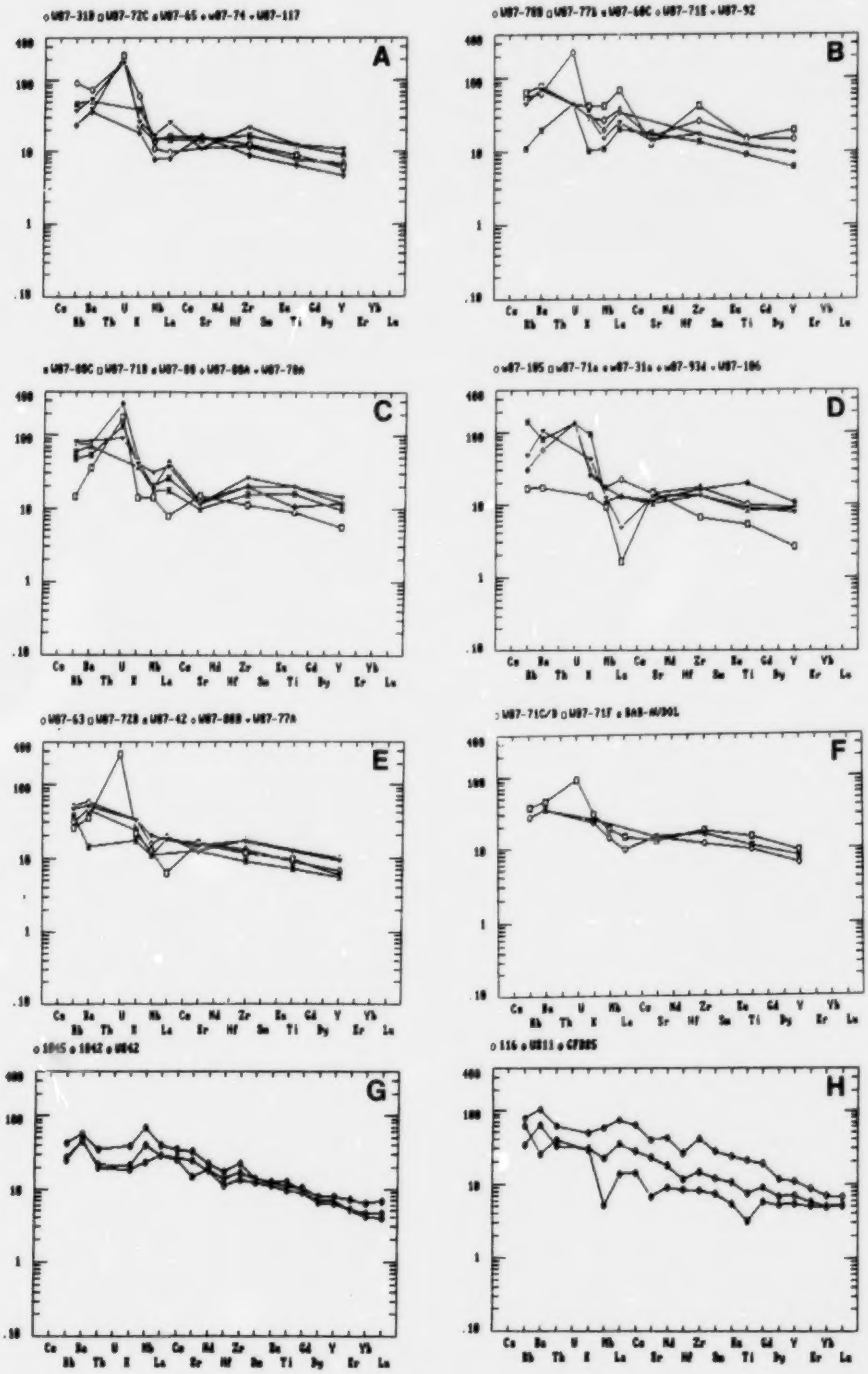

Figure 8. *A to h are spidergrams for Naskaupi sill diabase samples. The elemental concentrations are ratioed to the primitive mantle values of Taylor and McLennan (1981) and Jenner et al. (1987). Baragar's (1981) average value for the Naskaupi sills is sample bar-avdol. G and h are spidergrams of Thompson et al. (1983; 1984) continental flood basalts; g only resembles ocean-island basalts (note: the Thompson et al. (1983, 1984) data sets do not contain U concentrations).*

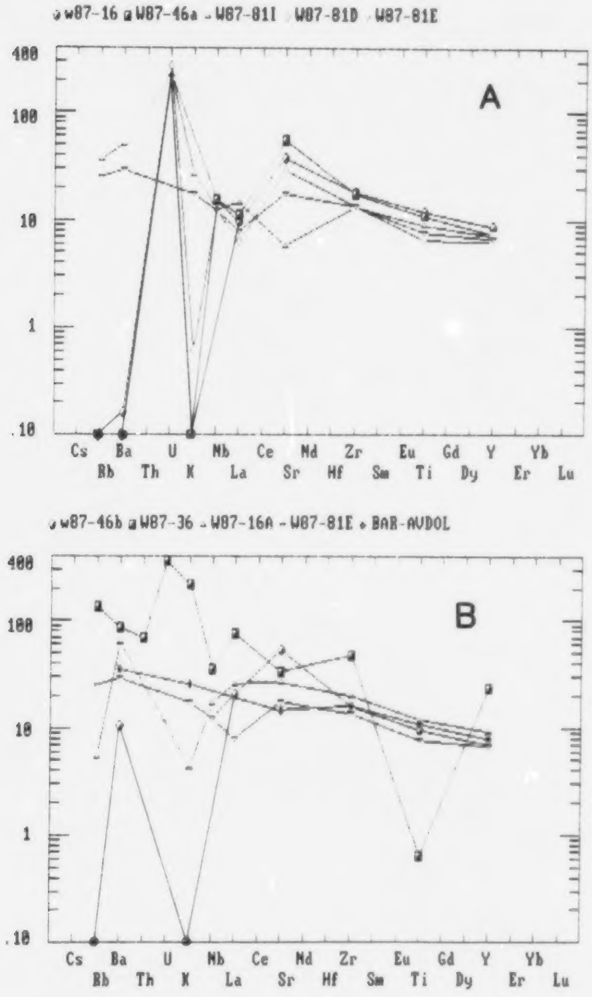

Figure 9a and b. *Spidergrams for mineralized diabase samples from the Naskaupi sills. The elemental concentrations are ratioed to the primitive mantle values of Taylor and McLennan (1981) and Jenner et al. (1987). Sample bar-avdol is Baragar's (1981) average value for the Naskaupi sills. A, represents those sill samples from copper occurrences and B, those from sills in proximity to copper occurrences.*

Sample W87-16 (Figure 8) is from a small island northwest of Adeline Island, wherein the host diabase is cut by quartz–carbonate veins and minor bornite and malachite (± chalcocite). Sample W87-46A is from the Seal Lake Main Showing, and sample W87-81D is from the most thoroughly mineralized trench at the Whiskey Lake (Mineral Occurrence 64) Showing. All three samples show extreme depletion in Rb, Ba and K and enrichment in Sr (i.e., the carbonate-rich ore fluids leached Rb, Ba and K from the diabase and added Sr).

Aside from the graphic depiction of ore fluid reaction, these common patterns are very significant because they show that the mineralizing processes were the same throughout the Seal Lake Group (i.e., W87-46A is from a showing generally agreed to be in the Salmon Lake Formation, whereas, W87-81D is definitely from stratigraphically higher mineralization). It has been suggested by Gandhi and Brown (1975) and Baragar (1981) that there were two periods, or styles, of copper mineralization in the Seal Lake Group *viz.*; a stratigraphically lower, epigenetic native copper style and a stratigraphically higher, syngenetic copper sulphide one. The geochemical data described here further support Wilton's (*this volume*) supposition that copper mineralization in the Seal Lake Group was probably the product of epigenetic ore fluids that were emplaced along structural weaknesses during a period of deformation

CONCLUSIONS

Based on these preliminary results, the Naskaupi sills do not appear to have significant potential for PGE concentrations. The geochemistry of the sills indicate that the parental magmas fractionated prior to intrusion and assimilated minor amounts of granulitic, or depleted, crustal material. Overall, the sills have geochemical signatures that are typical of continental flood basalts. Epigenetic copper-bearing hydrothermal fluids produced local, distinctive geochemical changes marked by depletion of Ba, K, and Rb.

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MIGUELS LAKE AREA (2D/12): AN UPDATE OF THE GEOLOGY

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ABSTRACT

New information on the geology of the Miguels Lake area has become available since it was originally mapped at the 1:50,000 scale in 1981. The Spruce Brook Formation has been dated paleontologically as Middle Ordovician (late Llanvirn to early Llandeilo). The subaerial Stony Lake volcanic rocks have been dated radiometrically at 423 ± 3 Ma. New field work in the northeast of the area has located the contact between the Silurian Botwood Group and granitic rocks of the Mount Peyton intrusive suite. New exposures of the fault zone between the Botwood Group and the Spruce Brook Formation have also been discovered; the exposures include altered ultramafic rocks.

INTRODUCTION

The Miguels Lake area, in central Newfoundland, includes the northern boundary between the Mount Cormack Subzone and the Dunnage Zone (Williams *et al.*, 1988) (Figure 1). It was mapped at a scale of 1:50,000 in 1981 and an open file map (Colman-Sadd and Russell, 1981) was released the same year. Shortly after the release of the map, paleontological dating of one of the main units was completed by R. Neuman (United States Geological Survey); this information and the revised interpretation that it necessitated were incorporated into papers by Colman-Sadd and Russell

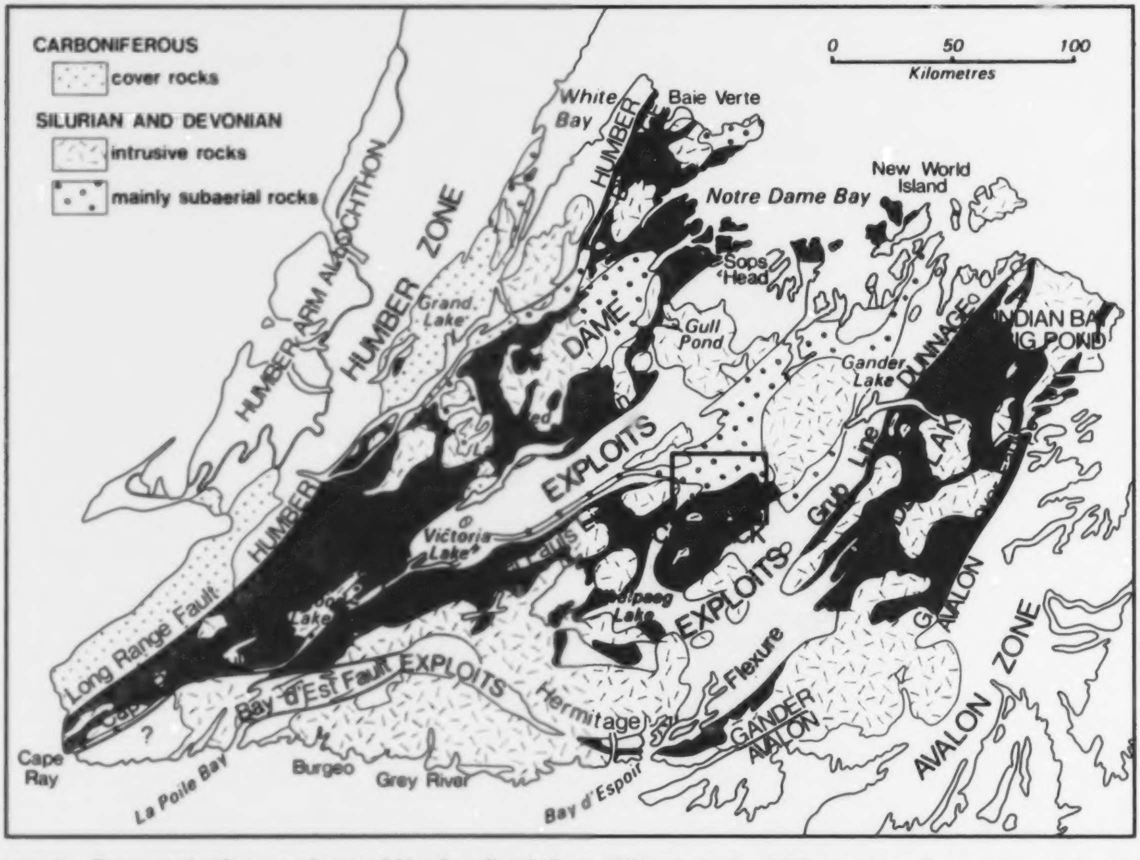
CARBONIFEROUS
cover rocks
SILURIAN AND DEVONIAN
intrusive rocks
mainly subaerial rocks
0
50
100
Kilometres
White Bay
Baie Verte
HUMBER
New World Island
Notre Dame Bay
HUMBER ARM ALLOCHTHON
HUMBER ZONE
Sops Head
Grand Lake
DAME
Gull Pond
INDIAN BAY
BIG POND
Gander Lake
DUNNAGE
EXPLOITS
Grub Line
Victoria Lake
Lake
Lake
Long Range Fault
Cape Ray
Bay d'Est Fault
EXPLOITS
Hermitage
Flexure
GANDER
AVALON
AVALON
AVALON ZONE
La Poile Bay
Burgeo
Grey River
Bay d'Espoir
?

Figure 1. *Zones and subzones of central Newfoundland (from Williams et al., 1988), showing the location of the Miguels Lake area.*

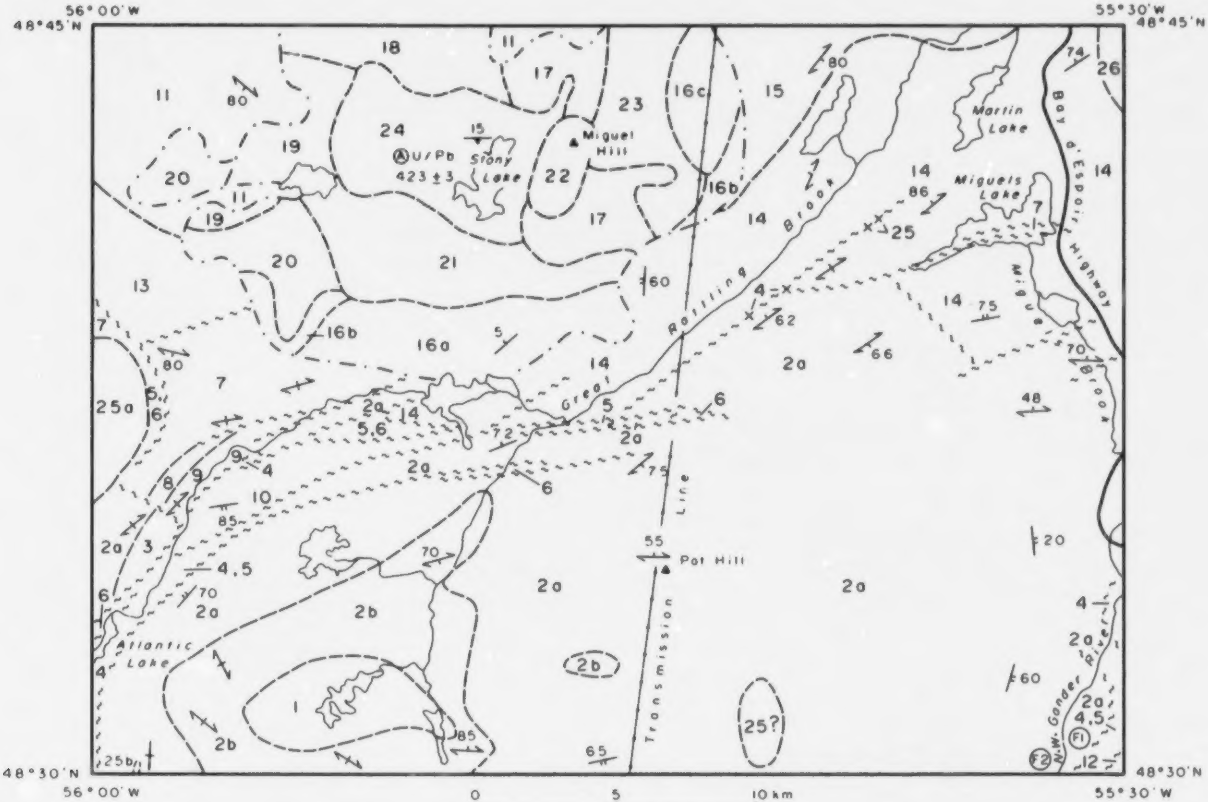

Figure 2. *Geological sketch map of the Miguels Lake area (after Colman-Sadd and Russell, 1988).*

(1982) and Colman-Sadd and Swinden (1984). Since then, one of the other main units has been radiometrically dated (Dunning *et al.*, 1988) and has been found to be substantially older than previously thought. As well as these revisions, which only affect the map legend, there have been changes in the map pattern (Figure 2) resulting from field work carried out during 1988, along a system of newly built logging roads. All of the new information is included in a revised version of the 1:50,000 map (Colman-Sadd and Russell, 1988) and is described in this paper.

PALEONTOLOGICAL DATING

Two fossil localities were found in limestone conglomerate in the southeast of the area during the original mapping. The conglomerate exposures are along strike from each other, and appear to form a horizon that is interbedded with shale and quartzite of the Spruce Brook Formation. However, actual contacts have never been observed, so this relationship is not completely certain. Brachiopods, in both the matrix and the pebbles of the conglomerate from the Northwest Gander River locality (F1, see Figure 2), were identified by R. Neuman, and the identifications were reported by Colman-Sadd and Swinden (1984). Neuman (*in* Colman-Sadd and Swinden, 1984) concluded that both faunas indicate an Ordovician age, ranging from the late Arenig to the early Caradoc, but most probably Llanvirn. The second locality (F2, see Figure 2), on a tributary of the Northwest Gander River, was sampled in 1985 and yielded a rich fauna of cephalopods, gastropods, trilobites and brachiopods. Boyce (1987) reported an age determination of late Llanvirn to early Llandeilo based on the cephalopods and trilobites, and Neuman (personal communication, 1987) produced a similar result from an examination of the brachiopods.

Colman-Sadd and Russell (1981) originally suggested a Silurian age for the Spruce Brook Formation, which the fossil data now clearly show to be incorrect. The Middle Ordovician age presented problems of tectonic interpretation because the Spruce Brook Formation is mainly of continental derivation, whereas surrounding rocks are oceanic or island arc in character, but are also thought to be Middle Ordovician. The conflict was resolved by considering the latter rocks to be allochthonous, as described by Colman-Sadd and Russell (1982) and Colman-Sadd and Swinden (1984).

RADIOMETRIC DATING

The Stony Lake volcanics were listed as 'Devonian or younger' by Colman-Sadd and Russell (1981), because they appear to lie on the Silurian Botwood Group with angular unconformity and to be posttectonic. However, Dunning *et*

LEGEND

DEVONIAN OR OLDER

26 MOUNT PEYTON INTRUSIVE SUITE: *porphyritic biotite granite*

25 *Granite:* 25a, *biotite–garnet granite of* OVERFLOW POND GRANITE; 25b, *muscovite–garnet pegmatitic granite*

SILURIAN

STONY LAKE VOLCANIC ROCKS (UNITS 16 TO 24)

24 *Dark-grey to black feldsparphyric rhyodacite; mainly pyroclastic, locally flow-banded*

23 *Dark-grey to dark-green feldsparphyric rhyodacitic lapilli tuff*

22 *Grey and purple flow-banded or brecciated rhyolite*

21 *Purple and pink rhyolite; mainly pyroclastic, locally flow-banded*

20 *Light purple, feldsparphyric rhyolite; mainly pyroclastic, locally flow-banded*

19 *Grey rhyolite; mainly lapilli tuff, locally flow-banded*

18 *Medium grained quartz and feldspar rhyolitic crystal tuff*

17 *Mainly buff, but locally green and purple, feldsparphyric, pyritiferous rhyolite; mainly flow-banded, but locally pyroclastic*

16 *Siltstone and sandstone:* 16a, *grey;* 16b, *red;* 16c, *grey and red containing blocks of flow-banded rhyolite and pyroclastic rocks*

BOTWOOD GROUP (UNITS 14 AND 15)

15 *Red sandstone and siltstone*

14 *Green-grey sandstone and siltstone with minor red siltstone beds*

13 ROGERSON LAKE CONGLOMERATE: *Unexposed, extrapolated from outcrop in Noel Paul's Brook area*

DUNNAGE ZONE

MIDDLE ORDOVICIAN AND OLDER

VICTORIA LAKE–BAIE D'ESPOIR GROUPS

12 NORTH STEADY POND FORMATION: *Unexposed, extrapolated from outcrop in Burnt Hill area*

11 TALLY POND FORMATION: *Interbedded felsic crystal tuff and lapilli tuff*

10 SALMON RIVER DAM FORMATION(?): *Interbedded purple-grey sandstone, siltstone and minor shale*

Pipestone Pond–Great Bend Complexes and other ophiolitic rocks

9 *Black shale interbedded with fine grained sandstone*

8 *Conglomerate, composed mainly of mafic volcanic clasts*

7 *Mafic pillow lava, breccia and tuff*

6 *Gabbro and minor trondhjemite*

5 *Pyroxenite*

4 *Peridotite and altered ultramafic rocks*

GANDER ZONE

ORDOVICIAN OR SILURIAN

3 *Strongly foliated granite*

MIDDLE ORDOVICIAN AND OLDER

2 SPRUCE BROOK FORMATION: 2a, *interbedded grey sandstone, siltstone and shale and minor conglomerate and limestone beds, metamorphosed in the greenschist and amphibolite facies;* 2b, *migmatite and granitoid gneiss thought to be derived from 2a.*

1 *Interbanded, migmatized, granitic, amphibolitic, semipelitic and psammitic gneisses*

SYMBOLS

Geological contact ⁄ ⁄ ⁄

Presumed unconformity ⁄ · ⁄ · ⁄

Fault ~ ~ ~ ~

Bedding, tops known (inclined) ⟋

Bedding, tops unknown (inclined, vertical) ⟋ ✕

Eutaxitic layering in volcanic rocks (inclined) ⟋

Cleavage (inclined, vertical) ⟋ ✕

Location of radiometric dating sample Ⓐ

Fossil localities (F1) (F2)

Newly discovered fault exposures mentioned in text X *al.* (1988) have reported a U–Pb zircon date of 423 ± 3 Ma from the volcanics, which places them at the top of the Wenlock stage of the Silurian (Haq and van Eysinga, 1987). Since the Botwood Group contains Wenlock fossils, the main deformation of this group and perhaps many of the older units in the area, is much more closely constrained than was previously thought. This conclusion is dependent on there actually being an angular unconformity between the Stony Lake volcanics and the Botwood Group, a relationship that has yet to be demonstrated.

The new date on the Stony Lake volcanics shows them to be approximately contemporaneous with the other main Silurian volcanic sequences in Newfoundland; in north-central Newfoundland, the Springdale Group has been dated at 429 $^{+6}_{-5}$ Ma (Chandler *et al.*, 1987) and in southwest Newfoundland the La Poile Group has yielded several ages between 426 and 419 Ma (Dunning *et al.*, 1988). The variable intensity of deformation in these other groups indicates that orogeny had not ended at the time that the Stony Lake volcanics were deposited, and the posttectonic appearance of the volcanics is therefore either fortuitous or suspect.

FIELD WORK 1988

Since the original mapping in 1981, logging operations by Abitibi-Price have resulted in the construction of a network of woods-roads in the northeast quadrant of the map area. These roads were investigated during 1988 for the new exposures created, and some of the problems raised by these new exposures were followed up by re-examination of several stream sections.

Mount Peyton Intrusive Suite

In the extreme northeast of the map area, logging has exposed an intrusive contact of granite into sandstone, siltstone and shale of the Botwood Group. The contact effects of the intrusion on the country rocks are very restricted. An exposure 1.5 km from the contact is similar in texture to the rest of the group, but has unusually rusty weathering, apparently caused by more pyrite than is normal for these rocks. This is possibly the result of fluid action related to the intrusion. An exposure of the Botwood Group, which is about 400 m from the nearest granite exposure, shows clear evidence of contact metamorphism and has been baked to form a black, uncleaved rock with a porcelaneous appearance. At the contact itself, green-grey to black sedimentary rock has been net-veined by fine grained, white granite containing white feldspar phenocrysts, 3 to 5 mm across (Plates 1 and 2). Scattered patches of coarser, pink pegmatite with tourmaline, occur at the edges of some of the veins. The main part of the granite is only exposed in one location within the map area, and even here it is close enough to the contact to contain abundant xenoliths of country rock. The granite is dark grey, medium grained, and contains biotite and phenocrysts of white feldspar. A granitic phase on the west side of the Mount Peyton intrusive suite was recognized by Kean and Mercer (1981) in the Grand Falls area (2D/13) and by Colman-Sadd (1982) in the West Gander Rivers area (2D/1 The new exposure in the Miguel's Lake area demonstrat the continuity of this phase, but still does not indicate t location of the contact with gabbro that outcrops in t northern part of the suite (Tomlin, 1982).

Plate 1. *Metasedimentary rocks of the Botwood Group, n veined by granite of the Mount Peyton intrusive suite.*

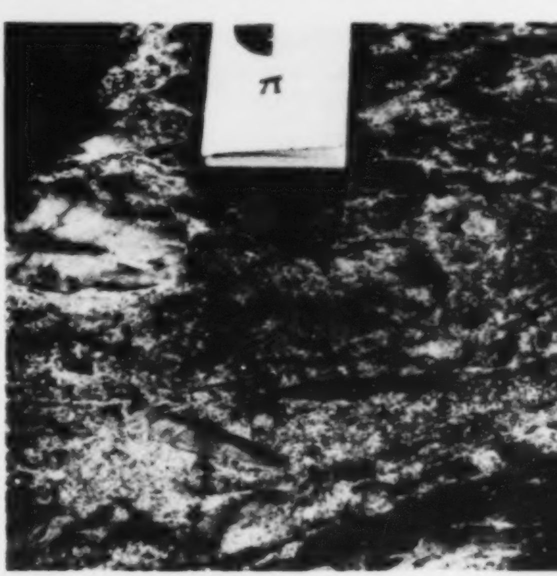

Plate 2. *Feldspar–porphyritic granite of the Mount Peyt intrusive suite, containing xenoliths of Botwood Gro metasedimentary rocks.*

Botwood Group

Most of the new exposures created since 1981 are in the more southerly unit of the Botwood Group (Unit 14) and consist of green-grey sandstone, siltstone and shale, with thin interbeds of maroon shale. Ripple marks are common and most beds show fine-scale cross-lamination (Plates 3 and 4); mud cracks occur in a few places. All the rock types in the unit are maroon on weathered surfaces. A single cleavage is present (Plate 5) and some exposures show tight folding.

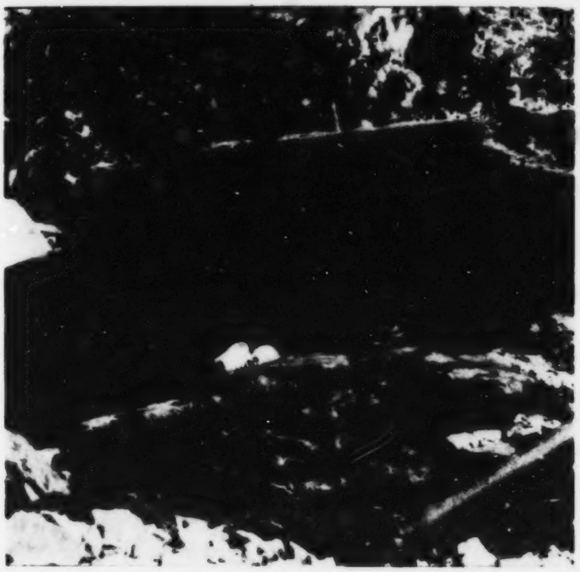

Plate 3. *Ripple marks on the surface of a Botwood Group sandstone bed.*

Plate 4. *Cross-laminations in a sandstone bed of the Botwood Group.*

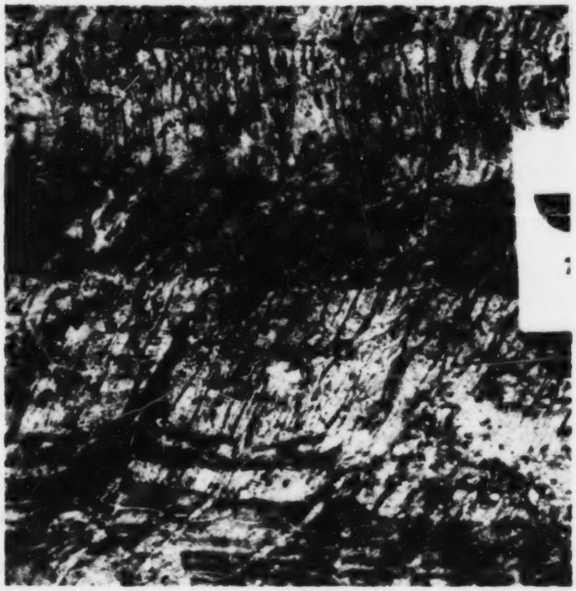

Plate 5. *Cleavage in the Botwood Group, showing refraction from shale to sandstone beds.*

Rocks exposed in Miguels Brook west of the Bay d'Espoir highway were briefly re-examined and are now provisionally included in the Botwood Group, pending further work in 1989.

Spruce Brook Formation

Interbedded white-weathering quartzite and dark-grey shale of the Spruce Brook Formation are newly exposed on woods-roads, west and south of Miguels Lake. In both cases, the rocks are at about the biotite grade of metamorphism and are distinctly more deformed than the Botwood Group. West of Miguels Lake, the exposures are small and consist principally of parallel-laminated shale with only a few thin quartzite beds; there appear to be two penetrative cleavages in the pelitic beds. South of Miguels Lake, the quartzite beds are up to 1-m thick, slightly boudined and have a fracture cleavage; the pelite beds are much thinner, strongly cleaved, and contain dismembered fragments of thin quartzite layers.

Boundary Faults

In previous publications, the fault zone assumed to separate the Botwood Group from the Spruce Brook Formation has been projected northeast, from pyroxenite and gabbro exposures near the confluence of North and South Great Rattling brooks, to the mylonitized mafic volcanics at Miguels Lake. This fault zone is now exposed on woods-roads west and northwest of Miguels Lake, and also has been found in a stream section where it was missed in the earlier mapping. Its actual location is about 2 km northwest of where it was predicted to be, and it coincides with and explains a prominent northeast-trending linear aeromagnetic anomaly (Geological Survey of Canada, 1968).

The exposure on a new road, about 5.5 km west of Miguels Lake and 1 km southeast of Great Rattling Brook, is marked by a prominent accumulation of red-weathering boulders formed of magnesite-rich altered ultramafic rock, net-veined by quartz. This rock type does not occur in the *in situ* exposure at this locality, which is instead formed of dark-green, mylonitized serpentinite schist. The schist contains sheared fragments, some of which appear to have been derived from mafic volcanic rock, and has numerous rootless isoclinal folds. A lineation plunges moderately toward the northeast and steep, northeast–southwest-striking CS fabrics indicate a sinistral sense of shear (Plate 6). The Botwood Group is exposed about 500 m to the west (a cross-strike distance of about 250 m) and the Spruce Brook Formation about 1.5 km to the southeast.

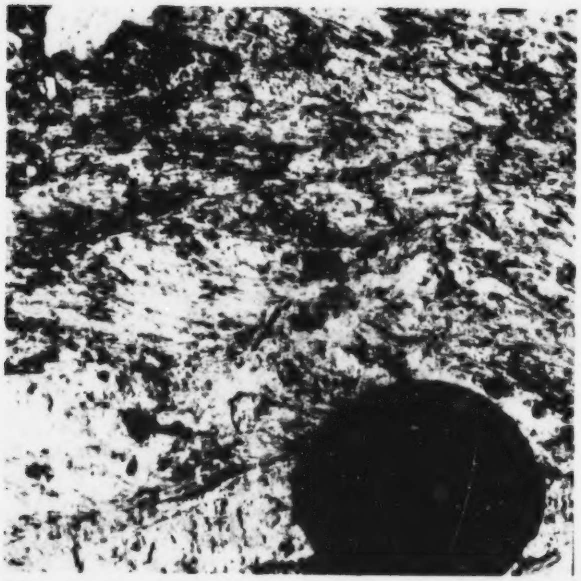

Plate 6. *Tectonic fabrics in serpentine schist, marking the fault zone exposed beside a woods-road, 5.5 km west of Miguels Lake.*

The boundary fault is better exposed 2 km to the southwest in a stream section. In the northern part of the stream, cleaved green-grey sandstone and shale of the Botwood Group dip southeast. At its southeast contact, the dip reverses through a tight syncline to a northwesterly direction and the sediments overlie massive magnesite-rich altered ultramafic rock along a sharp irregular boundary. The contact is assumed to be a fault rather than an unconformity because of its sheared nature, both here and along strike, and because there is no variation in the composition or grainsize of the sedimentary rocks. The massive magnesite at the contact is only about 2-m wide and gives way southeast to about 5 m of cleaved magnesite with boudins of the massive material, and then passes into black pyritic slate. The slate occupies a zone about 500-m wide and terminates at an exposure of bedded quartzite and pelite of the Spruce Brook Formation, which occurs at the north end of the first pond above Great Rattling Brook. The slate is in fact a mélange in which the cleavage is frequently crosscut by shear zones. Bedding is unrecognizable and quartzite occurs as boudins or rootless isoclinal fold closures. Tectonized fragments of magnesite or serpentine-rich altered ultramafic rock are also common in certain parts of the zone and are generally 10- to 30-cm across; an exception is a large block of serpentinite, several metres wide at about the midpoint in the mélange unit. Variably deformed quartz veins occur sporadically throughout the zone.

The coincidence of the observed fault zone with a linear aeromagnetic anomaly allows the fault to be traced where exposure is poor or ambiguous. The fault clearly extends into the area northwest of Miguels Lake where the Botwood Group is exposed on both sides of it, providing further evidence that the contact between the Botwood Group and the ultramafic rocks is tectonic. In this area, two exposures of foliated granite have been found along the projected fault trace and, since their lithology and state of deformation are anomalous in the context of the surrounding Botwood Group, it is assumed that they have been tectonically introduced along the fault zone. The granite is grey or pink, equigranular and inhomogeneously deformed, and has accessory black tourmaline or amphibole. Typically, it has an anastomosing fabric with sericite coating the cleavage surfaces. The more easterly of the two exposures contains a lens of massive pyrite, 10 by 30 cm, aligned along the cleavage and is cut by posttectonic vuggy quartz veins. The fault zone that has now been defined southeast of Great Rattling Brook is not aligned with the fault previously mapped on the islands in Miguels Lake (Colman-Sadd and Russell, 1981). It is uncertain how the two are related, but it is considered most likely that the Miguels Lake fault is a splay off the main system, and this is how it is depicted by Colman-Sadd and Russell (1988). Both faults are presumed or seen to juxtapose the Botwood Group against the Spruce Brook Formation and both extend into the main outcrop area of the Botwood Group; both are marked by strongly tectonized slithers of presumed ophiolitic rocks.

CONCLUSIONS

The Miguels Lake area contains an unusual variety of rock units ranging from successions typical of both the Gander and Dunnage zones, to Silurian overlap sequences that apparently span the main deformation in this part of Newfoundland. A full understanding of the relationships in the area is some way off and can be expected to contribute considerably to our knowledge of Newfoundland geology and the history of Appalachian orogeny. Perhaps the most critical unresolved problem is the nature of the contact between the Botwood Group and the Stony Lake volcanics. Paleontological and radiometric dating have been essential ingredients in the mapping, both for constructing an accurate map legend and for highlighting the regional implications. The follow-up field work during 1988 has confirmed the rough predictions of the original mapping and has greatly improved the map's accuracy. It has also defined a major fault zone, containing deformed ultramafic rocks, which is analogous to faults along the Gander River Ultrabasic Belt and the Baie Verte Line. The comparison suggests high potential for gold mineralization (Tuach *et al.*, 1988).

ACKNOWLEDGMENTS

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STRATIGRAPHY AND SEDIMENTOLOGY OF THE PORT AU PORT GROUP (MIDDLE TO UPPER CAMBRIAN), WESTERN NEWFOUNDLAND: PRELIMINARY RESULTS

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ABSTRACT

Preliminary results of detailed stratigraphical and sedimentological analyses of Middle and Upper Cambrian platform strata in western Newfoundland, indicate a possible alternative to current regional correlations. Relative inboard–outboard facies relationships between measured sections, may dictate further revisions in Cambrian platform paleogeographical models, and may also indicate significant tectonic juxtaposition of terrains previously interpreted as autochthonous.

INTRODUCTION

During 1988, a project was initiated to study the stratigraphy and sedimentology of autochthonous platformal Cambrian strata in western Newfoundland. Stratigraphic sections were measured, and sedimentological field relationships were recorded for the carbonates and minor siliciclastics of the March Point, Petit Jardin and Berry Head formations (Middle to Upper Cambrian) in the areas of the Port au Port Peninsula, South East Arm of Bonne Bay, and Goose Arm of Bay of Islands (Figure 1). At these locations, the units are as yet unmapped at the formation level (1:50,000). This study is in its preliminary stages and the results presented are still tentative and subject to re-interpretation, pending further field surveys during the four-year lifetime of the project.

Analyses of data collected will enhance the resolution of regional correlations of the Cambrian carbonate lithostratigraphic units and allow statistical validation of possible stratigraphic bundling (i.e., grouping and correlation of small-scale carbonate cycle bundles) in these rocks. With this data, the paleogeography of the Cambrian platform and nature of the ancient continental margin can be better reconstructed, and, also, will clearly constrain the large-scale tectonic relationships of the autochthonous and allochthonous strata in western Newfoundland. In addition, if the statistical analysis does reveal stratal bundling at more than one scale, then the deciphering of such packaging would provide an efficient correlation tool for base-metal exploration in the western Newfoundland Cambro–Ordovician strata.

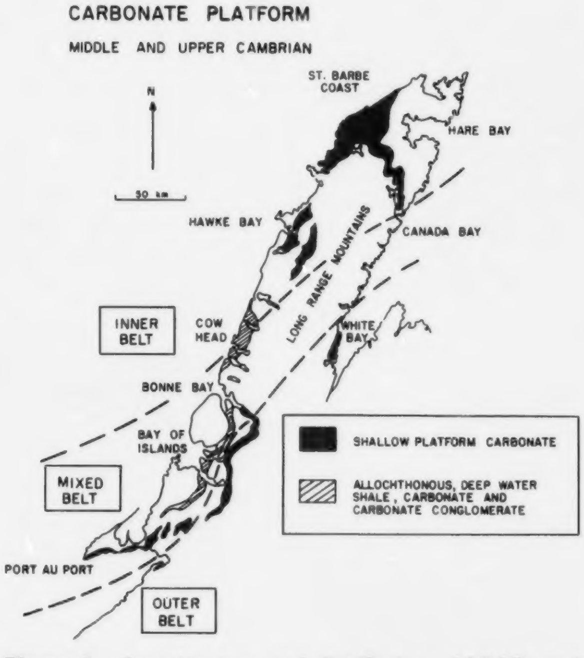

Figure 1. *Location map and distribution of Middle and Upper Cambrian strata in western Newfoundland; major facies belts delineated by previous workers include 1) an inner mixed siliciclastic-carbonate belt, 2) an intermediate belt in which inner facies and oolite are mixed, and 3) an outer belt of stacked oolite sand shoals; (from James et al., 1983).*

GEOLOGICAL SETTING

Autochthonous Cambrian strata of the Labrador and Port au Port groups record sedimentation on the ancestral North American (Laurentian) continental margin, adjacent to the Iapetus Ocean (Williams and Stevens, 1974). This ancient trailing margin, evolved from a rift–drift setting to a thermally stable passive margin in the late Proterozoic to

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This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

earliest Cambrian (James and Stevens, 1982). Williams and Hiscott (1987) have suggested that the stratigraphic signal of the rift–drift transition occurs possibly within, but probably no later than, the Bradore Formation (lower Labrador Group), and the subsequent Cambrian to Lower Ordovician deposition represents sedimentation on a rigid lithosphere, characteristic of a stable passive margin experiencing a linear subsidence rate (James *et al.*, *in press).*

The Lower Cambrian Forteau Formation (Bonnia-Ollenellus Zone), represents deposition of limestone, shale, siltstone and minor sandstone on a pre-platform, Lower Cambrian mixed siliciclastic–carbonate shelf (James *et al.*, 1988). Terrigenous clastics and minor carbonates of the overlying Hawke Bay Formation (*Bonnia-Olenellus* to *Bathyuriscus-Elrathina* and *Ehmaniella zones;* Knight and Boyce, 1987) represent a possible pan-Iapetus drop in sea-level, and the influx of littoral siliciclastic facies across the Forteau shelf (Palmer and James, 1979). Overlying Middle to Upper Cambrian March Point, Petit Jardin and Berry Head formations consist of limestones, dolostones and minor fine-grained terrigenous siliciclastics (Knight, 1977, 1980; Levesque, 1977) (Figure 2).

Grand Cycles

Stratigraphic superposition of the members of the March Point, Petit Jardin and Berry Head formations have been interpreted by Chow and James (1987a) to constitute at least three Grand Cycles (cf. Aitken, 1966, 1978) in western Newfoundland. Grand Cycles consist of *lower* shaly and *upper* carbonate half cycles (Figure 2). Alternate shale and carbonate half cycles are interpreted to result from lateral shifts in two adjacent sedimentary mega-environments, in response to eustatic sea-level fluctuations. In western Newfoundland, the two mega-environments consist of, 1) a subtidal oolite shoal complex associated with back-shoal intertidal–supratidal flats (carbonate half cycle), which lie relatively outboard of, and adjacent to, 2) a fine-grained, mixed, carbonate–siliciclastic tidal-flat system (shaly half cycle) (Chow, 1986). Repetitive vertical stacking of carbonate and shaly half cycles suggested to Chow and James (1987a) that the paleogeographical character of the Middle to Upper Cambrian carbonate platform was persistent, if not static, through time. They interpreted the present regional distribution of strata as a reflection of the former relative positions of major facies belts on the Cambrian platform (Figures 1 and 3).

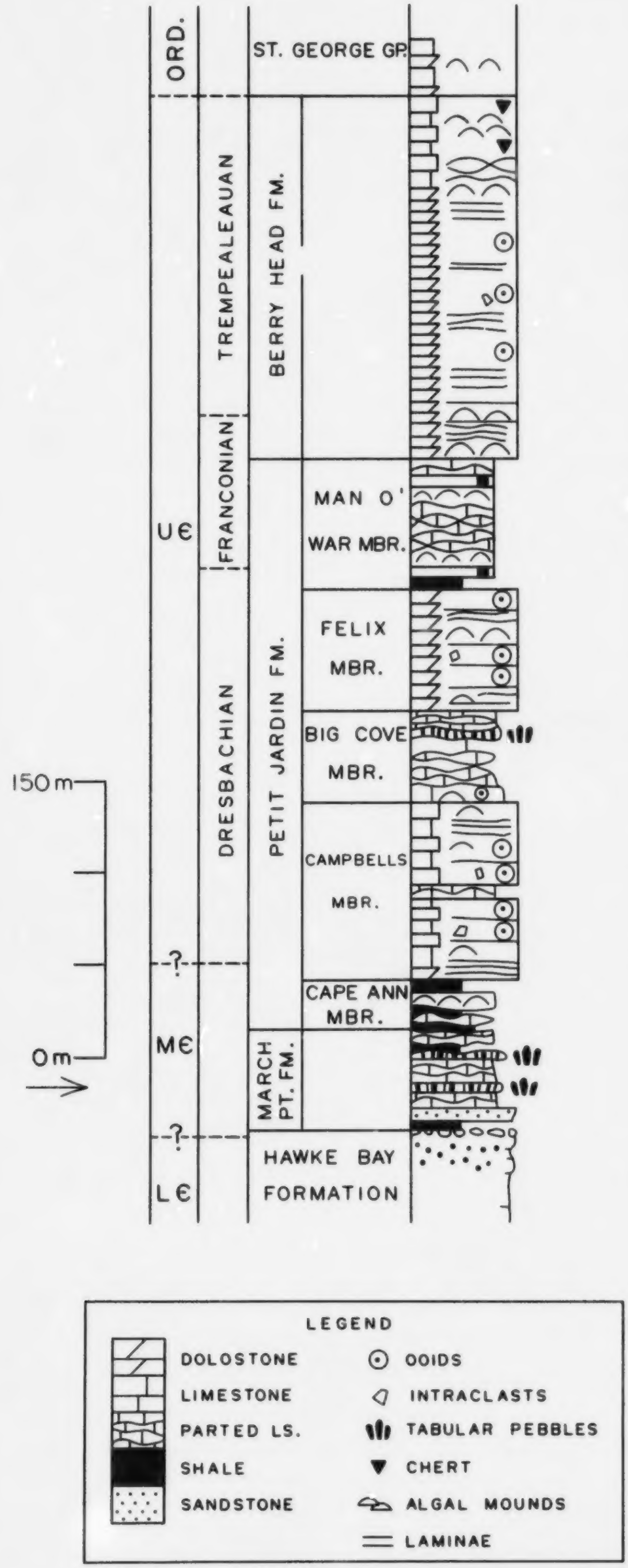
ORD.
ST. GEORGE GP.
TREMPEALEAUAN
BERRY HEAD FM.
FRANCONIAN
U€
MAN O' WAR MBR.
DRESBACHIAN
PETIT JARDIN FM.
FELIX MBR.
BIG COVE MBR.
CAMPBELLS MBR.
CAPE ANN MBR.
150 m
-?-
M€
MARCH PT. FM.
0 m
-?-
HAWKE BAY FORMATION
L€
LEGEND
DOLOSTONE
LIMESTONE
PARTED LS.
SHALE
SANDSTONE
OOIDS
INTRACLASTS
TABULAR PEBBLES
CHERT
ALGAL MOUNDS
LAMINAE

Figure 2. *Stratigraphic column of Middle and Upper Cambrian strata in western Newfoundland; from Chow and James (1987a); Grand cycles are expressed as couplets; March Point-Cape Ann/Campbells; Big Cove/Felix; Man O'War/Berry Head.*

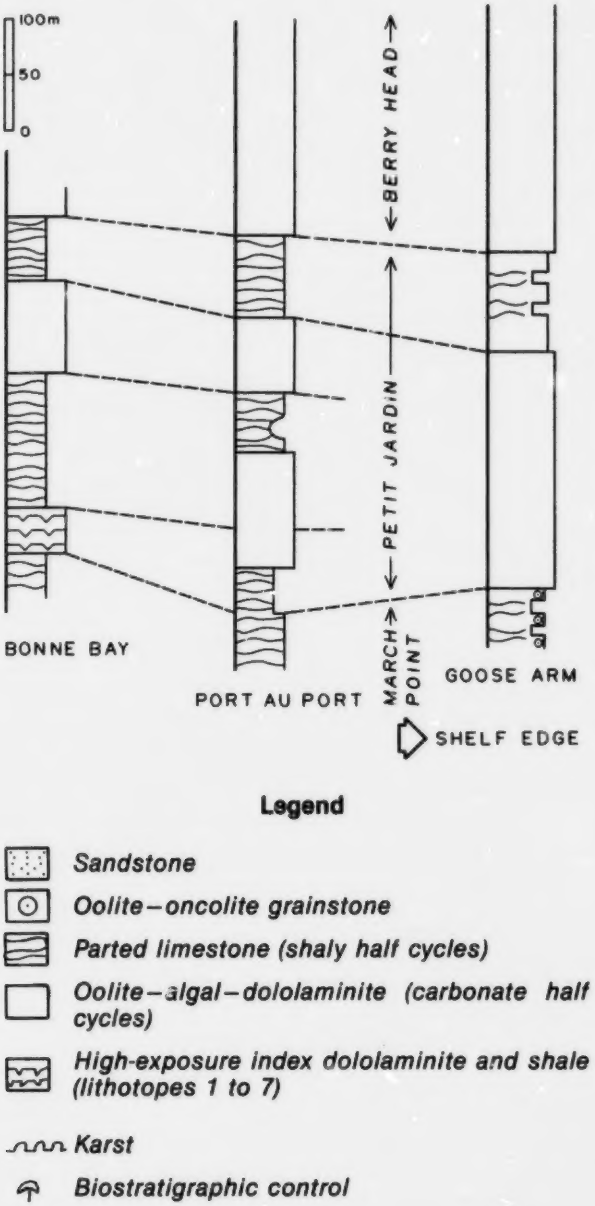
100m
50
0
BERRY HEAD
PETIT JARDIN
MARCH POINT
BONNE BAY
PORT AU PORT
GOOSE ARM
SHELF EDGE
Legend
Sandstone
Oolite–oncolite grainstone
Parted limestone (shaly half cycles)
Oolite–algal–dololaminite (carbonate half cycles)
High-exposure index dololaminite and shale (lithotopes 1 to 7)
Karst
Biostratigraphic control

Figure 3. *Previous lithostratigraphic correlations of Middle and Upper Cambrian strata for sections in this investigation; modified from Chow (1986).*

PRELIMINARY RESULTS

During the 1988 survey season, work focussed on describing the Middle and Upper Cambrian strata, specifically the detailed sedimentology and stratigraphy of the Grand Cycles of the March Point, Petit Jardin and Berry Head formations.

Sedimentology

Carbonate Half Cycles. Sediments of the Middle and Upper Cambrian carbonate half cycles (Campbells member, Felix member, and Berry Head formation), are currently categorized into three basic lithotopes: grey oolite (subtidal), brown oolite (intertidal) and dololaminite (supratidal) (Chow and James, 1987b). Metre-scale or smaller algal mounds (e.g., LLH stromatolites or thrombolites) and minor shale, punctuate the oolite–dololaminite sequences. These strata are here subdivided into fourteen basic lithotopes (Table 1). Metre-scale packages of these lithotopes comprise the carbonate half cycles at each locality. However, this is not to imply that metre-scale packages are composed of all fourteen lithotopes, indeed, rarely do more than five or six lithotopes occur in any given package. Composition and internal organization of these metre-scale packages is complex, and a predictive model, which would determine lithotope composition and proportion within a given package, has as yet, not been developed. Similarly, the stacking order of lithotopes is not uniform throughout a given section or from area to area. Further analysis is needed to clearly indicate, whether these lithotopes are organized in statistically predictable packages, and/or whether different lithotopes, as delineated here, can occupy the same stratigraphic position within packages at separate localities or at stratigraphically different positions in a larger scale package.

It appears that deposition occurred above storm wave-base, and possibly above normal wave-base, for each of the carbonate half cycles. This is suggested by the following evidence; predominance of exposure surfaces, lack of large-scale and hummocky bedforms in sediments of the appropriate grain sizes, and the monotonous thickness of the beds (<1 m) and lithotope packages (<2 m).

With the exception of the Goose Arm section, in which several-metre-thick oolite and oncolite beds occur in the March Point Formation, the beds of oolite in carbonate half cycles are <1-m thick, and generally lack large-scale cross-stratification. This does not support Chow's (1986) interpretation, of the grey oolite as a subtidal mobile shoal facies fringed by a back shoal tidal flat and represented by the brown oolite and dololaminite lithotopes. However, if Chow and James' (1987b) interpretation that these oolite beds represent *in situ* sedimentation is accepted, then the thin oolites cannot be explained as washover deposits. Rather, they represent deposits of short-lived, shallow marine environments as suggested by the occurrence of intercalated algal mounds.

These Cambrian sediments have many characteristics in common with algal–oolite shoreface to foreshore environments of Great Salt Lake, Utah, and Shark Bay, Western Australia, including type and distribution of fauna and sediment, and scale of bedforms (Eardley, 1938, 1966; Logan, *et al.*, 1970, 1974). The physical characteristics of

Table 1. *Lithotopes delineated in this investigation for the carbonate half cycles of Cambrian Grand Cycles, western Newfoundland*

LITHOTOPE	DESCRIPTION	RELATIVE EXPOSURE INDEX	OBSERVED IN GRADATIONAL CONTACT WITH
1	Shale, buff to grey, dolomitic, mudcracks common	High	2-7
2	Very argillaceous dolostone, orange to buff, mudcracks common	High	1,3-7
3	Argillaceous dolostone, beige, blocky	High	1,2,4-7
4	Dolostone, minor argillaceous, buff to beige, blocky	High	1-3,5-7
5	Dolostone, buff, argillaceous, laminated, mudcracked	High	1-4,6-8
6	Dolostone, buff, blocky, laminated to massive	High	1-5,7,8,10,11
7	Dolostone, buff, blocky, laminated to massive, very intraclastic	High	1-6,8,10,11
8	Oolite, cross-bedded, minor to abundant dololutite flasers and clasts	Interm	5,6,7,9-12
9	Oolite, clean, grey to buff	Interm	8,10-14
10	Dololutite-parted limestone, grainy (peloidal to oolitic)	Interm	5-9,11
11	Dololutite-parted limestone, syneresis, flaser- to lenticular-bedded	Interm	5-9,10,12
12	Argillaceous dolomite-parted limestone, flaser- to lenticular bedded	Interm	8,9,12,13,14
13	Shale-parted limestone, wave ripples, cruziana ichnofacies	Low	9,12,14
14	Shale, dark grey, fossiliferous calcareous and oolite lenses rare	Low	9,12,13

these modern, shallow, low-gradient, epeiric environments, which have a restricted fauna, may provide a better analogy than do the Bahamian-type oolite environments for the sediments of the Middle and Upper Cambrian in the study area, where the oolite shoals are characterized by large-scale accretionary bedding (Ball, 1967, Halley *et al.*, 1983), lack of small-scale cyclicity and subaerial exposure, and a scarcity of stromatolites.

Shaly Half Cycles. The March Point Formation and Cape Ann, Big Cove, and Man O'War members of the Petit Jardin Formation on the Port au Port Peninsula and Bonne Bay, are characteristically shaly and lack significant oolites. These units consist of shale- and/or dololutite-parted limestones, associated with flaser to lenticular bedding that show small-scale current and wave-ripple cross-stratification. Intraclastic, flat pebble conglomerates are common, and algal bioherms, some stacked up to 3-m high, punctuate these successions. These sediments are interpreted by Chow (1986) as mixed carbonate–siliciclastic tidal-flat deposits, grouped in shallowing-upward flaser- to lenticular-bedded metre-scale packages. This tidal-flat system is envisaged as inboard of the oolite shoal–back-shoal flat system of the carbonate half cycles.

In order to integrate these parted limestone units into the overall stratigraphy, a paleoenvironmental interpretation must be made that is consistent with the interpretation of the depositional environments for the carbonate half cycles. In this report, the shaly half cycle sediments are re-interpreted as mixed, fine-grained (peloidal mud to fine sand) carbonate and siliciclastic sediments deposited below normal, but above storm wave-base. Significantly, the flaser- to lenticular-bedded packages of Chow (1986) could not be delineated in the field, due to the lateral variability and stratigraphically gradational characteristics of the flaser- and lenticular-parted limestones. On the basis of preliminary data presented here, it is apparent that this facies occupied a paleogeographic position outboard to, and gradational with, the algal–oolite shoreface to foreshore environments of the oolitic carbonate half cycles, and inboard to the thick subtidal oolite shoal sequences at Goose Arm. This reconstruction places the flaser- to lenticular-parted limestones in the lee of a subtidal oolite shoal belt of the subtidal to very low intertidal zone (*Cruziana* ichnofacies; Frey and Pemberton, 1984). Further work will determine the efficacy of this preliminary interpretation.

STRATIGRAPHY

Stratigraphic correlations of Middle and Upper Cambrian rocks, in the sections measured, have been previously proposed by Chow and James (1987a). They attributed the generation of Newfoundland Grand Cycles to global sea-level fluctuations, through comparisons made with other North American Cambrian Grand Cycles. However, Grand Cycles, as defined by shaly and carbonate half cycle couplets, are correlated with difficulty throughout the Great Northern Peninsula because the number of couplets expressed at different localities is variable (Figure 3). Chow and James (1987a) attributed this variation to the relative positions of stratigraphic sections with respect to major facies belts on the Cambrian platform.

An alternative stratigraphic framework and approximate gross correlation is presented as a preliminary result of this investigation. Within the constraints of known biostratigraphical control, the thick sequence of shales and mud-cracked dololaminites at Bonne Bay (lithotopes 1 to 7;

Table 1) are possible time-equivalents to the Campbells member of the Petit Jardin Formation on the Port au Port Peninsula (Figure 4). Therefore, it is possible that the Bonne Bay sediments represent significant inboard deposition relative to the sections on the Port au Port Peninsula. In contrast, the Goose Arm section experienced the most outboard depositional conditions, at least during deposition of the March Point Formation as evidenced by the thick beds of oolite and oncolite grainstone, which dominate the sequence. Correlation of strata above and below the Campbells member can be made to maintain this inboard to outboard relationship of the various sections, such that interpreted time-equivalent interval thickness and average bedset thickness increase paleoshoreward. This re-interpretation is preliminary, and at this point in the investigation we do not favour any one interpretation, rather, alternatives are being tested in order to develop an overall stratigraphic model.

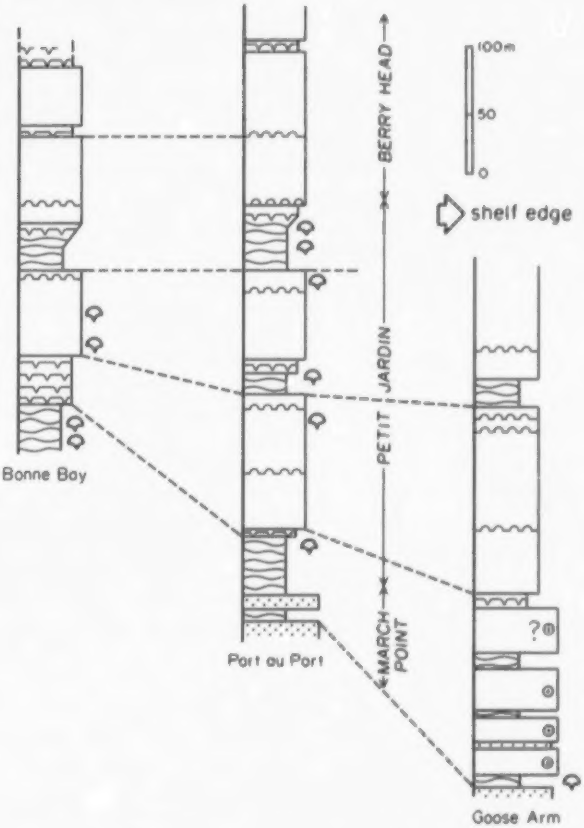

Figure 4. *Lithostratigraphic interpretation of the Middle and Upper Cambrian strata; see Figure 3 for legend.*

Care was taken to correlate gross stratal intervals, bounded above and below, by disconformities, i.e., major karst surfaces. These major surfaces, typically with an overlying quartzose carbonate, occur near the tops of each carbonate half cycle, and are recognized as regional events by Chow (1986). Quartz is patchy in temporal and spatial distribution, and is neither present on every major karst surface nor ubiquitous over the study area for a given correlatable karst surface. In this way, the preliminary correlations are consistent with sequence stratigraphic principles.

Further work, particularly measurement of sections at Little Coney Arm of White Bay and along the St. Barbe Coast, will help clarify these proposed paleogeographic and stratigraphic relationships. Statistical work will reveal if any small-scale rhythms exist in these strata, which would be critical to refining and increasing the resolution of these gross correlations. It is of fundamental importance to understand the relative proximal to distal positioning of these sections because small-scale rhythms would be expressed differently at each location as a result of differences including accommodation, supply of sediment and length and frequency of exposure/nondeposition.

CONCLUSIONS

This report is a preliminary revision of previous stratigraphical, sedimentological and paleogeographical interpretations of the Middle and Upper Cambrian carbonate strata of western Newfoundland. Of primary significance to exploration in Cambro–Ordovician autochthonous strata throughout the study area, is the recognition that Cambrian sections exposed in the areas of study may represent disparate inboard and outboard platform sedimentation. Documentation of paleoenvironmental gradients expressed by such an inboard/outboard disparity, is critical to understanding regional facies distributions, which in turn would control the potential distribution of facies-dependent base-metal mineralization.

Re-interpretation of paleogeographic positions of the major facies belts in the Newfoundland Cambrian Grand Cycles has implications for the models for the generation of Cambrian Grand Cycles in general (Aitken, 1978; Chow and James, 1987a). That is, Grand Cycle superpositioning of carbonate half cycles over shaly half cycles is re-interpreted here as the result of progradation of most proximal over most distal facies, rather than the reverse. Further work is required before the causal mechanism for Grand Cycle development can be determined. It must also be noted that statistical testing of the preliminary interpretation presented here must be sought in the fine-scale organization of stratal packages within the Grand Cycles.

Present distribution of the measured stratigraphic sections has strong implications for tectonism associated with the closing of the Iapetus Ocean. Present geographic arrangement of the most inboard (Bonne Bay section), intermediate (Port au Port) and most outboard (Goose Arm) strata, suggests significant tectonic juxtaposition of these sections may have occurred, and thus revision of paleogeographic models currently in use for Middle and Upper Cambrian rocks may be in order. Further implications as to the amount of transport experienced by imbricate thrust slices in which these strata occur must be evaluated, as well as the possibility of northeast–southwest oriented tectonic telescoping of these terrains.

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SEDIMENTOLOGICAL STUDIES IN THE EASTPORT BASIN: NOTES ON THE PETROGRAPHY OF THE LATE PRECAMBRIAN CONNECTING POINT GROUP AND PROVENANCE IMPLICATIONS

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ABSTRACT

The sedimentary rocks of the Connecting Point Group are volcaniclastic lithic arenites, and are compositionally similiar to the pyroclastic rocks of the upper part of the underlying Love Cove Group. The lower 1500 m of the succession consists of felsic volcanic and metamorphic detritus and a subordinate andesitic component. Andesitic clasts represent the dominant lithic component of sandstones, beginning at a pebbly mudstone (mixtite) and continuing upward for over 1500 m. However, in this upper interval, the fine-grained tuffs, and the matrix and blocks of tuff within the mixtite, suggest that felsic volcanism was continuous, and that only the proportion of andesitic to felsic volcanism changed at this stratigraphic level.

The petrographic data suggest that the source terrain of the volcaniclastic sediments was a mature island-arc complex, developed on a continental crust and/or island-arc crust that included medium- to high-grade metamorphic rocks.

The ratio of quartz to pyroclastic detritus, which is indicative of sediment maturity, declines upward. This is interpreted to reflect, 1) the widening of the rifted arc basin so that narrow, shallow-water zones of sediment reworking were isolated from the zones of sediment remobilisation on the volcaniclastic apron, and 2) the large volume of pyroclastic material generated in the arc, overwhelmed the depositional basin, and caused slope instability and rapid sediment remobilisation. An episode of basin extension, previously correlated with the formation of mixtite and intrusion of mafic dykes and plutons, appears to coincide with the increase in the proportion of andesitic detritus in the sediments of the upper part of the Connnecting Point Group.

INTRODUCTION

The late Precambrian Connecting Point Group is a marine, volcaniclastic sequence at least 3500 m thick (Hayes, 1948). Preliminary lithostratigraphical and sedimentological surveys by Knight and O'Brien (1988), indicate that the lower 2400 m are the product of sedimentation in a system of submarine aprons, adjacent to a volcanic arc. The volcanic arc is considered to be represented by the Love Cove Group, equivalent age volcanic sequences and related plutons that lie west of the western Avalon Zone.

There are two major objectives of this petrographic study: 1) to determine the provenance of the detritus in the upper Love Cove and Connecting Point groups, and thus advance our understanding of the geotectonic setting of its sedimentation, and, 2) to evaluate aspects of emplacement of the volcaniclastic sediments, such as the deposits of gravity-driven sediment flows.

The diagenesis, metamorphism and deformation of the discussed sediment deposits have not been examined in detail. The sediments have been regionally metamorphosed to a greenschist facies. Deformation and accompanying metamorphism affect the sequence most severely in the vertical to inverted, eastward-facing limb of the Willis Reach syncline (Figure 1), where there is substantial bedding-parallel flattening of the strata. This affects the rocks principally in the Long Reach–Pretty Island and Hail Island–Coal Island sections. Gently folded and dipping strata, east of the area around Eastport and Salvage, are almost pristine in their preservation of volcanic textures.

STRATIGRAPHY AND LITHOFACIES

Pyroclastic rocks and epiclastic sedimentary rocks of the upper part of the Love Cove Group and the overlying Connecting Point Group, form a conformable succession in the Eastport basin (O'Brien and Knight, 1988). Shales, and fine to coarse pyroclastic rocks and minor tuffaceous turbidites, characterize the uppermost part of the Love Cove Group. An assemblage of ten lithofacies combine in different proportions to constitute the six lithostratigraphic units that make up the Connecting Point Group (Knight & O'Brien, 1988). Five of these units (C-1 to C-5) are shown in the graphic logs on Figure 2; the sixth unit (C-6) has not been logged in detail and is thus not shown on the figure.

The lower 1750 m of the Connecting Point Group form an upward-coarsening succession composed of three units.

This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

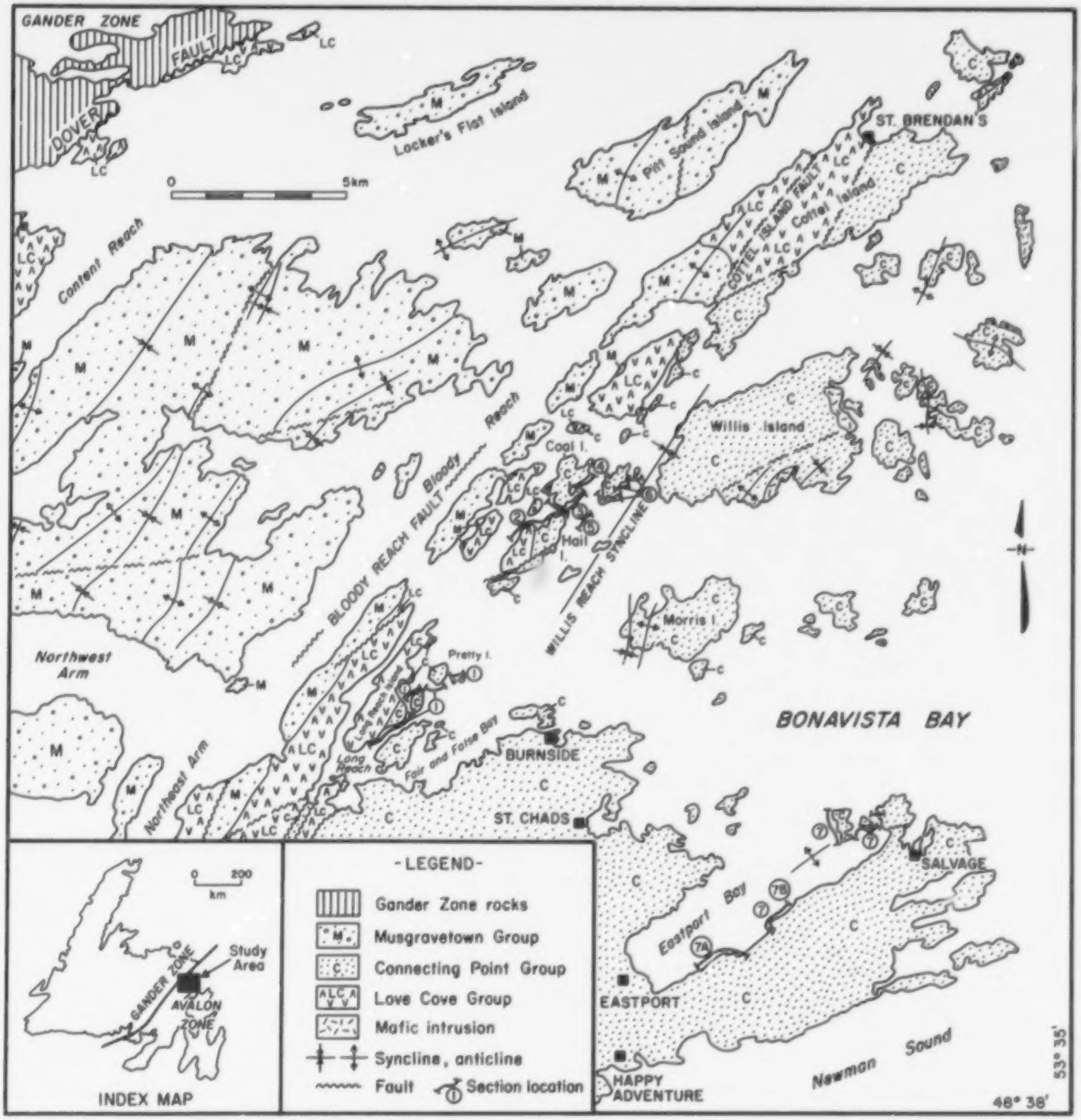
GANDER ZONE
DOVER FAULT
Locker's Flat Island
Pitt Sound Island
ST. BRENDAN'S
COTTEL ISLAND FAULT
Cottel Island
0
5km
Content Reach
Bloody Reach
BLOODY REACH FAULT
Coal I.
Willis Island
Hail I.
WILLIS REACH SYNCLINE
Morris I.
Pretty I.
Northwest Arm
Long Reach Island
Long Reach
Fair and False Bay
BURNSIDE
BONAVISTA BAY
Northeast Arm
ST. CHADS
SALVAGE
Eastport Bay
EASTPORT
HAPPY ADVENTURE
Newman Sound
N
53° 35'
48° 38'
M
C
LC
-LEGEND-
Gander Zone rocks
Musgravetown Group
Connecting Point Group
Love Cove Group
Mafic intrusion
Syncline, anticline
Fault
Section location
INDEX MAP
0
200
km
Study Area
GANDER ZONE
AVALON ZONE

Figure 1. *Geological map of the Eastport area of Bonavista Bay showing the location of stratigraphic sections and sampling; modified from O'Brien (1987), Knight and O'Brien (1988) and O'Brien and Knight (1988).*

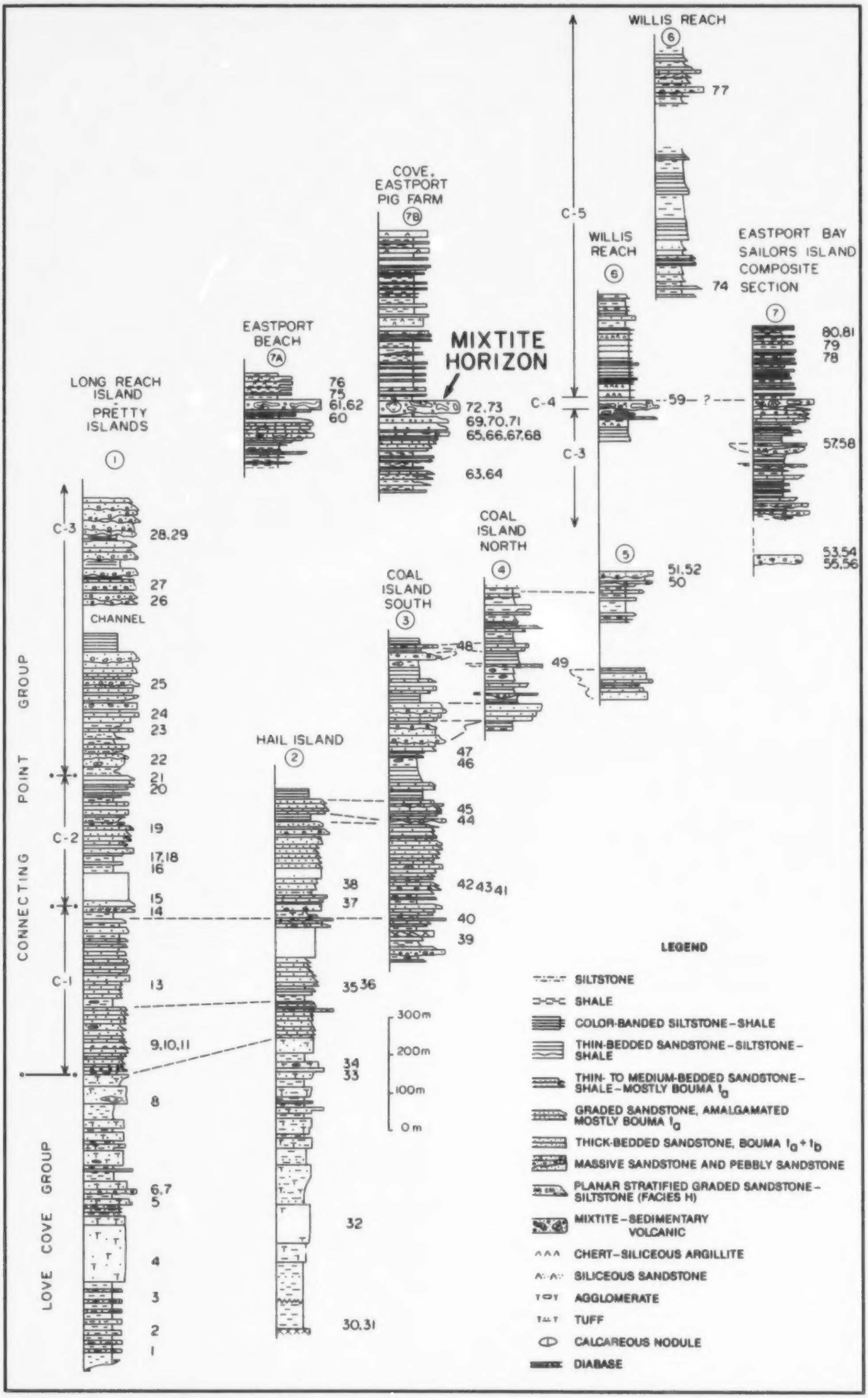
WILLIS REACH
⑥
77
74
C-5
COVE, EASTPORT PIG FARM
(7B)
MIXTITE HORIZON
72,73
69,70,71
65,66,67,68
63,64
WILLIS REACH
⑥
EASTPORT BAY SAILORS ISLAND COMPOSITE SECTION
⑦
80,81
79
78
C-4
59
?
57,58
C-3
53,54
55,56
EASTPORT BEACH
(7A)
76
75
61,62
60
LONG REACH ISLAND
PRETTY ISLANDS
①
C-3
28,29
27
26
CHANNEL
25
24
23
22
21
20
19
C-2
17,18
16
15
14
C-1
13
9,10,11
8
6,7
5
4
3
2
1
CONNECTING POINT GROUP
LOVE COVE GROUP
COAL ISLAND NORTH
④
⑤
51,52
50
49
COAL ISLAND SOUTH
③
48
47
46
45
44
42 43 41
40
39
HAIL ISLAND
②
38
37
35 36
34
33
32
30,31
300m
200m
100m
0 m
LEGEND
SILTSTONE
SHALE
COLOR-BANDED SILTSTONE-SHALE
THIN-BEDDED SANDSTONE-SILTSTONE-SHALE
THIN- TO MEDIUM-BEDDED SANDSTONE-SHALE-MOSTLY BOUMA t_a
GRADED SANDSTONE, AMALGAMATED MOSTLY BOUMA t_a
THICK-BEDDED SANDSTONE, BOUMA $t_a + t_b$
MASSIVE SANDSTONE AND PEBBLY SANDSTONE
PLANAR STRATIFIED GRADED SANDSTONE-SILTSTONE (FACIES H)
MIXTITE-SEDIMENTARY VOLCANIC
CHERT-SILICEOUS ARGILLITE
SILICEOUS SANDSTONE
AGGLOMERATE
TUFF
CALCAREOUS NODULE
DIABASE

Figure 2. *Graphic logs of sections through the upper Love Cove Group and the overlying Connecting Point Group showing stratigraphic location of sample numbers; C 1 to 5 lithostratigraphic units and sedimentary logs, modified from Knight and O'Brien (1988, Figures 3 and 4).*

The basal unit (Unit C-1; Figure 2) comprises 400 m of thin-bedded, green–grey and buff, tuffaceous sandstone and siltstone, black shale and rare fine grained pyroclastic rocks and coarse-grained epiclastic conglomerate and sandstone. Thicker bedded, massive sandstones also occur, but bed-thickness or grain-size trends are not obvious in the unit. The succeeding unit (Unit C-2; Figure 2) is 370-m thick and consists of thickening- and coarsening-upward sequences of parallel-bedded, thin- to thick-bedded, graded, very fine- to very coarse-grained sandstone, rare pebbly sandstone and shale. The upper unit (Unit C-3; Figure 2) comprises 1000 m of thinly bedded and laminated sandstone, siltstone and shale, enclosing large lenticular, tens of metres thick, fining-upward bodies of massive sandstone and pebbly sandstone. Intervals of parallel-bedded turbidites and black shales also occur. A regional pebbly mudstone (mixtite) unit, 11- to 17-m thick (Unit C-4; Figure 2) lies above the third unit. It contains variably sized rip-up clasts and rafts of intrabasinal sediments, which are deformed, as well as pebbles and boulders of volcanic and plutonic origin. It is in turn overlain by a 500-m-thick shale-dominated sequence (Unit C-5; Figure 2), which is also characterized by thin sandstone and siltstone beds, cherts and slump deposits. Several thick sandstone beds having planar stratification and water escape structures occur in the unit immediately above the mixtite. The section that has been studied so far is completed by a succession (Unit C-6; not shown on Figure 2) over 1000-m thick, of thin to thick-bedded, turbiditic sandstones and shales organized in repetitive coarsening- and thickening-upward packages.

The sandstones of the Connecting Point Group have the characteristics of high-density turbidites deposited in both distal and proximal settings of the basin (Knight and O'Brien, 1988). The organization of the sediments up through the section indicates that deposition occurred in a variety of deep-water submarine fan settings, including low-efficiency fans during the earlier part of the basin's history, and as channeled fans having extensive levée complexes, as the basin matured. The accretion of a very thick sequence of thin-bedded, fine-grained sediment and local evidence of slumping and mass-flow transport, suggest the progradation of the depositional slope, prior to an important episode of basin extension. This phase of extension coincided with the deposition of the mixtite, which is a regional olistostrome, and includes an intrusion of a suite of small plutons and dykes of mafic composition. Post-extension basin development is marked by progradation of submarine fans.

PETROGRAPHY

Methodology

Eighty-one thin sections, ranging from tuff to small pebble conglomerate, were examined to determine the primary composition of the deposits and their sedimentary, textural and structural characteristics. Thirty-five samples of 300 points per thin section were counted in order to determine proportions of their detrital components. Metamorphism and deformation have modified the sedimentary and petrographic signatures of the rocks in question. This has often precluded reliable distinguishing of some detrital constituents, such as metamorphic quartz, from fragments of recrystallised felsic groundmass; it has also inhibited distinction between intrabasinal tuff clasts and fragments of aphanitic felsic groundmass.

The proportions of various lithic fragments were determined to analyse their abundance and significance. The matrix component was point counted on a separate channel and thin sections were stained in order to identify K-feldspar.

The point-counting technique followed the approach of Zuffa (1980, 1985; see also Gazzi, 1966; Dickinson, 1970; Ingersoll *et al.*, 1984). The results, combined with detailed petrographic examination of the clasts, serve as the data base for evaluation of the provenance of the Connecting Point Group. Geochemical data is presently not available to precisely classify the volcanic clasts. However, on the basis of the petrography, the volcanic clasts are subdivided into two general categories, 1) rhyolite–dacite and, 2) andesite–basalt; the latter group is represented dominantly by andesitic clasts.

Tuffs and Tuffites

Tuff and tuffite horizons in the upper 1000 m of the succession help elucidate the nature of the pyroclastic components in the sandstones described below (e.g., see samples 71 to 73 and 75; Figure 2). In the tuff beds, shards (Plate 1) and pumice fragments are typically replaced by quartz and feldspar (probably albite) (Plates 1, 3 and 6). Chlorite, calcite and prehnite less commonly replace glass. Throughout most of the succession in the western limb of the Willis Reach syncline, felsic glass is apparently rare,

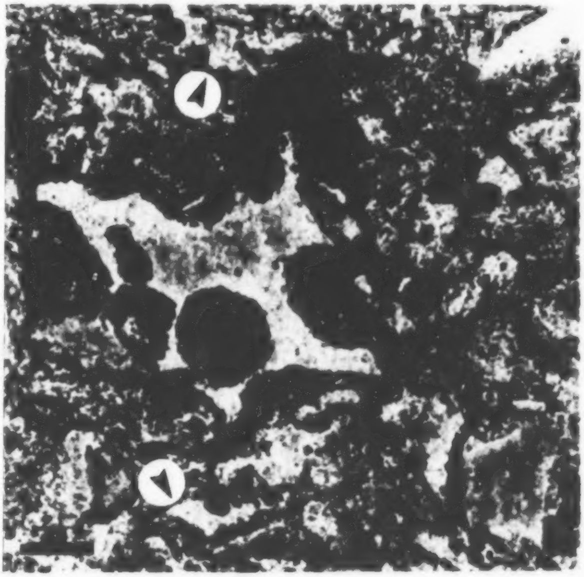

Plate 1. *Fragment of felsic pumice from a bed of crystal tuff; glass is altered to albite and chlorite and the unresolved matrix to chlorite and prehnite; note shards (arrows) and plagioclase pyroclasts (lower right corner); plane-polarized light photomicrograph of sample 71; scale bar = 0.05 mm.*

although its identification is complicated by considerable deformation. It is probable that a substantial proportion of vitric material has been dissolved during metamorphism- and deformation-related pressure solution (cf. Dimroth and Demarke, 1978).

Detrital Components of Sandstones (Connecting Point Group)

Lithic fragments

Rhyolite–dacite. Felsic volcaniclastic detritus in the entire sequence studied (Figure 2) is mostly of reworked, intrabasinal pyroclastic[1] origin, namely lithic and crystal tuff. Both pyroclastic and epiclastic fragments of cryptocrystalline quartz, and a mixture of quartz and feldspar, form the rhyolite–dacite lithic component of the sand fraction (Plates 2 and 3). Locally, clasts having quartz and plagioclase phenocryst, flow banding and fragments of spherulites, indicate that the observed 'chert-like' grains are fragments of aphanitic volcanic groundmass. The groundmass is dominated by cryptocrystalline, commonly spherulitic silica (Plate 2), which is locally recrystallised to mega-quartz, but also includes some K-feldspar. Idiomorphic areas are either irregular or linear, the latter defining flow banding.

Felsic volcanic sand grains in coarse-grained sandstones and conglomerates characteristically show porphyritic and glomeroporphyritic textures. Phenocrysts of quartz, up to 4 mm in size, are euhedral, embayed or considerably corroded having round outlines (Plate 2). Locally, biotite occurs as an accessory mineral.

Rare vesicular lithic particles have amygdules of quartz, chlorite, epidote and sphene. Felsic fragments exhibiting perlitic cracks occur in one sandstone sample (sample 26, Figure 2; Plate 4).

Andesite. Andesitic detritus is represented by a mixture of angular to well-rounded clasts of pyroclastic and epiclastic origin, ranging in size from sand to cobbles. It occurs throughout the entire succession of the Connecting Point Group, but conspicuously dominates the lithic fragments in the mixtite, as well as the deposits just below it, and also for 1500 m above it (Figure 2; Table 1).

Andesitic clasts are commonly porphyritic and glomeroporphyritic (Plate 5) containing embayed plagioclase and rarely augite phenocrysts (Plate 6) up to 1.2 mm long. The andesitic fragments range from holocrystalline to hypocrystalline in which the glass is replaced mostly by chlorite, often in association with subordinate sphene and epidote. Quartz, feldspar (albite?), calcite and prehnite occur as less common forms of glass alteration. Apart from devitrified glass, hypocrystalline andesitic clasts are characterized by vesicles and skeletal plagioclase (Plates 5, 7 and 8), arranged in trachytic and variolitic textures. Trachytic and variolitic textures are also common in holocrystalline andesitic grains.

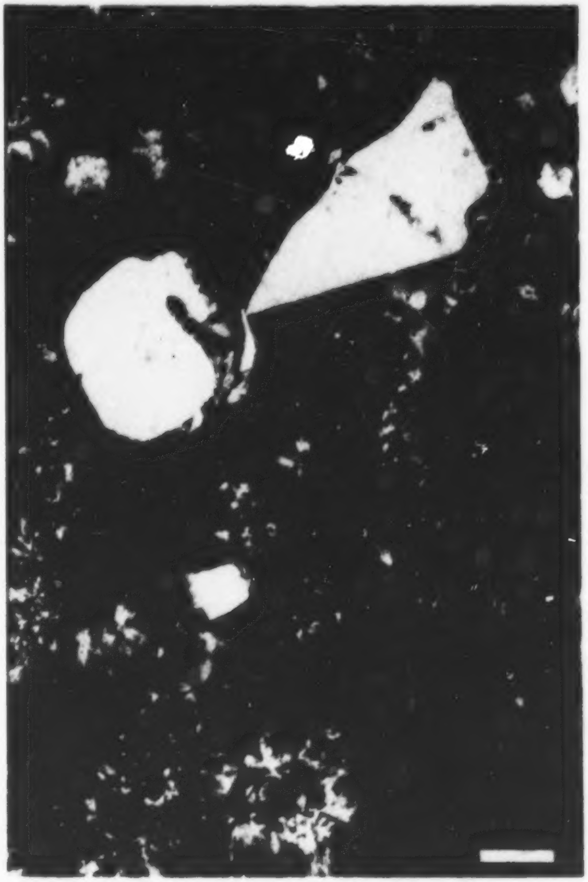

Plate 2. *A cobble-size epiclast of probable rhyodacitic composition; note the embayed quartz and feldspar phenocrysts and the spherulitic groundmass; cross-polarized light photomicrograph; scale bar = 0.25 mm.*

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[1] In this report, the following definitions are adopted: **pyroclasts** are particles expelled through volcanic vents without reference to the causes of eruption or origin of the particles; **epiclasts** are produced by weathering and erosion of the volcanic rocks. Reworking and recycling of unconsolidated pyroclastic debris by water or wind *does not* transform pyroclasts into epiclastic fragments (see Fisher and Schmincke, 1984, p. 89.).

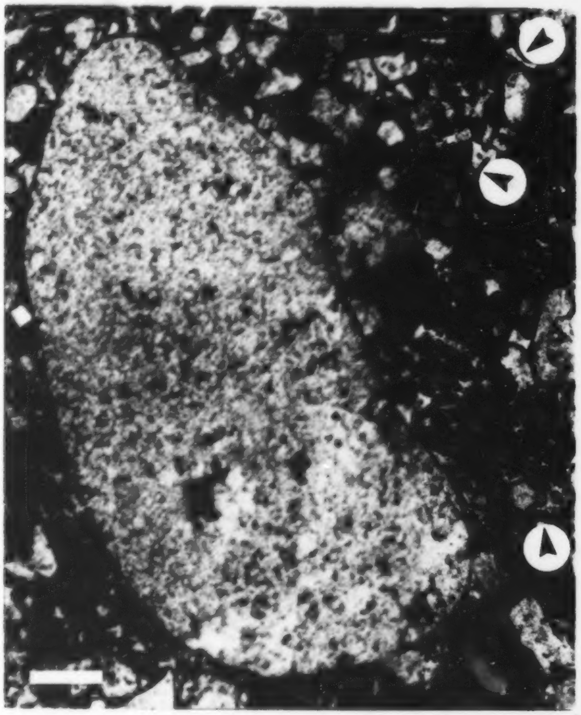

Plate 3. *Well rounded felsic epiclast surrounded predominantly by pyroclastic grains; arrows point to pumice fragments and shards; cross-polarized light photomicrograph of sample 72; scale bar = 0.25 mm.*

Pumice fragments (Plates 8 and 9) include hyaline types that appear to form a gradation of vesicular varieties derived from hypocrystalline andesite. Pumice fragments have irregular, jagged to multi-lobate shapes characteristic of lapilli (Plates 9 and 10). A subordinate proportion of rounded pumice fragments also occur. Ovoid to spherical vesicles are mostly filled with chlorite spherulites, commonly in association with quartz, feldspar and sphene. Calcite, clinozoisite and prehnite are less common in the amygdules. Shards of equivalent intermediate glass are rare, probably because they are less resistant to alteration compared to silicic glass (Fisher and Schmincke, 1984).

Schists. Schist fragments, as distinct from foliated volcanic tuff and groundmass altered to fine white mica and chlorite, occur in minor or trace amounts in the sand fraction throughout the studied sequence. Schist fragments, which are most abundant in a tuffaceous sandstone (sample 17, Figure 2 and Table 1; Plate 11), are represented by aggregates of muscovite ± biotite ± quartz ± garnet ± chlorite. Trace fragments of epidote-bearing siliceous schist have been identified in sample 54.

Intrusive rocks. Fragments of diorite (Plate 12) and microdiorite are a minor or trace constituent in the andesite-rich part of the succession. Microdiorite fragments are also

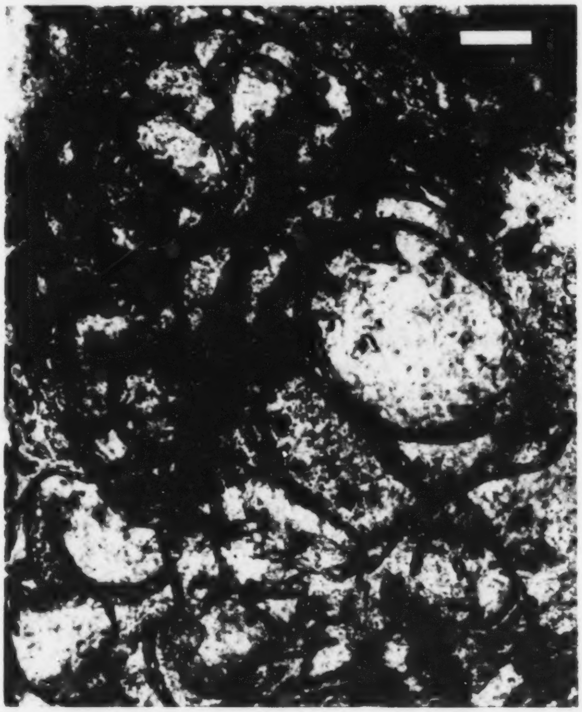

Plate 4. *Perlitic cracks in felsic sand grain; plane-polarized light photomicrograph of sample 63; scale bar = 0.05 mm.*

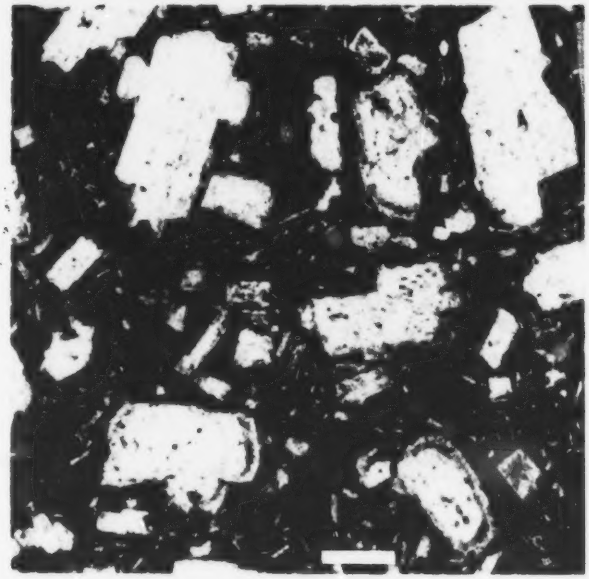

Plate 5. *Fragment of porphyritic andesite; large plagioclase phenocrysts showing zoning are embedded in a matrix of skeletal plagioclase crystals and cryptocrystalline groundmass that has been altered mainly to chlorite; plane-polarized light photomicrograph of sample 68; scale bar = 0.25 mm.*

Table 1: Mineralogy of detrital constituents in the deposits of the Love Cove Group (samples 1-8 and 30-34), and Connecting Point Group (the remaining samples)

Sample	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20
Quartz (monocrystalline)	22.0	.	.	33.7	-	.	6.3	22.7	25.1	3.3	29.0	23.8	30.5	.	.	-	20.0	.	26.3	16.8
Quartz (polycrystalline)	12.7	.	.	20.3	-	-	2.0	13.3	5.0	1.3	8.9	11.4	5.7	.	.	-	6.0	.	9.3	15.0
K-feldspar	1.0	-	.	6.3	-	-	1.3	6.3	2.3	-	5.3	1.7	6.5	-	-	-	3.0	-	8.7	1.5
Plagioclase	9.3	.	.	3.6	-	.	82.3	10.0	28.4	93.7	13.6	27.5	6.5	.	.	-	6.0	.	13.7	1.8
Lithic fragments																				
Andesite	-	?	.	-	.	-	1.7	0.3	3.6	-	1.4	1.0	-	.	.	-	1.0	.	-	1.5
Rhyolite–dacite	43.7	?	.	27.3	.	.	5.7	42.7	31.1	-	32.9	24.2	37.0	.	.	.	50.5	.	34.0	42.5
Shards–pumice	-	.	-	-	-	-	-	-	-	-	-	3.3	-	-	-	-	-	.	tr	-
Tuff–tuffite	-	-	-	-	-	-	-	-	-	-	tr	tr	-	-	-	-	-	-	-	tr
Schist	1.0	-	-	0.6	.	-	-	1.7	-	-	-	0.7	-	-	-	.	8.5	.	-	tr
Other lithic fragments	4.3	-	-	[2,1]7.7	-	-	[3]tr	3.0	[1]3.3	[3]1.6	[1]8.2	10.5	-	[3]	-	3.5	7.3	[3,2]17.1	-	[3,2]17.1
Accessory minerals																				
Zircon	tr	-	-	-	-	-	-	tr	-	tr	tr	-	tr	-	.	-	tr	-	-	tr
Sphene	tr	-	-	-	-	-	-	-	tr	-	-	tr	-	-	-	-	-	-	tr	tr
Garnet	tr	-	-	tr	-	-	tr	tr	tr	tr	tr	-	tr	.	.	-	tr	.	tr	tr
Epidote	tr	-	-	-	-	-	-	-	-	-	tr	tr	tr	-	-	-	tr	-	-	tr
Allanite	-	-	-	-	-	-	-	-	-	-	-	tr	-	-	-	-	-	.	-	-
Tourmaline	-	-	-	-	-	-	-	-	tr	-	-	-	-	-	-	-	-	-	-	-
Clinopyroxene	-	-	-	tr	-	-	-	-	-	-	-	tr	-	-	-	-	-	-	tr	-
Orthopyroxene	tr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Muscovite	-	-	-	-	-	-	-	-	-	-	-	tr	-	-	-	-	-	-	tr	-
Biotite	-	-	-	.	-	-	-	-	tr	-	tr	tr	-	-	-	-	1.0	-	-	-
Chlorite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	tr	-
Ore	2.3	-	-	-	-	-	-	-	-	-	-	0.7	-	-	-	-	-	-	-	-
Apatite	0.6	-	-	-	-	-	-	tr	tr	-	tr	-	tr	.	-	-	tr	.	tr	0.9
Matrix of the whole rock	76.4	X	X	X	X	X	44.9	67.0	81.5	39.0	86.6	80.6	83.8	X	X	X	83.5	X	79.7	69.7
Approximate grain size	cg	tuff cgl	cgl	mg	cgl	mg	cg	mg	cg	cg	cg	cg	fg	cgl	cgl	cgl	fg	mg	mg	cg

present; - absent; * – undifferentiated volcanic fragments; ? – identification uncertain due to recrystallization and deformation; X – not established; tr – trace; fg – fine-grained sandstone(ss.); mg – medium-grained ss.; cg – coarse-grained ss.; cgl – conglomerate; [1] – granite; [2] – undifferentiated intrusive rocks; [3] – microdiorite/diorite; [4] – trachyte; [5] – quartz arenite; [6] – quartzite.

Table 1: *(Continued)*

Sample	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	39	40	41	42
Quartz (monocrystalline)	10.5	.	13.0	5.3	11.3	17.0	11.7	.	16.7	.	.	.	.	.	.	.	.	.	.	.	.	.
Quartz (polycrystalline)	5.5	.	5.3	6.7	10.7	7.3	1.7	-	3.0	.	-	-	-	-	.	.	.	.	.	.	.	.
K-feldspar	2.0	-	3.3	1.0	4.3	3.7	1.0	-	-	-	-	.	-	-	.	.	.	.	-	-	.	.
Plagioclase	7.5	.	33.3	3.0	14.3	31.7	36.3	.	33.7	.	.	.	-	-	.	.	.	.	.	.	.	.
Lithic fragments																						
Andesite	-	-	7.0	-	4.7	7.0	7.3	.	8.7	.	.	-	-	-	.	-	.	.	.	.	-	-
Rhyolite–dacite	74.5	.	18.3	76.7	40.0	18.0	38.3	.	34.7	.	.	?	.	.	.	.	.	.	.	.	.	.
Shards–pumice	-	.	-	-	-	-	-	.	-	-	.	?	-	-	-	-	-	-	-	-	.	-
Tuff–tuffite	-	.	-	-	-	-	-	.	-	-	.	-	-	-	-	-	-	-	-	.	-	-
Schist	-	.	5.0	1.0	5.0	-	-	-	-		-	-	-	.	.	.	-	.	.	-	-	-
Other lithic fragments	-	-	12.3	5.0	7.8	^{3}15.0	^{3}2.7	^{4}	2.3	-	^{6}	-	-	-	^{1}		^{1}	-	^{2}	-	-	-
Accessory minerals																						
Zircon	-	-	tr	-	tr	tr	-	-	tr	.	-	.	-	-	-	.	-	.	.	-	.	.
Sphene	-	.	-	-	-	tr	-	-	tr	-	-	-	-	-	.	-	-	-	-	-	-	-
Garnet	-	-	tr	tr	tr	tr	tr	-	tr	.	.	-	-	-	.	.	-	.	.	.	.	-
Epidote	-	-	tr	-	tr	-	-	-	-	-	.	-	-	-	.	.	-	-	-	-	-	-
Allanite	-	-	tr	-	tr	-	-	-	tr	-	-	-	-	-	-	-	-	-	-	-	-	.
Tourmaline	-	-	-	-	tr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	.	-
Clinopyroxene	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Orthopyroxene	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Muscovite	-	-	tr	-	-	-	-	-	tr	-	-	-	-	-	-	-	-	-	-	.	-	-
Biotite	-	-	-	tr	tr	-	-	-	-	-	-	-	-	-	.	-	-	-	-	-	-	-
Chlorite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Ore	-	-	tr	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Apatite	-	-	tr	tr	tr	tr	tr	-	tr	.	-	-	-	-	.	.	-	.	-	.	.	.
Matrix of the whole rock	5.6	X	48.9	44.0	66.3	47.7	24.8	X	50.1	X	X	X	X	X	X	X	X	X	X	X	X	X
Approximate grain size	cgl	cg	cg	cg	mg	cgl	cg	cgl	cgl	cgl	cgl	cgl	cgl	cgl	fg	mg	cg	cg	cgl	cgl	mg	cg

. present; - absent; *—undifferentiated volcanic fragments; ?—identification uncertain due to recrystallization and deformation; X—not established; tr—trace; fg—fine-grained sandstone(ss.); mg—medium-grained ss.; cg—coarse-grained ss.; cgl—conglomerate; ^{1}—granite; ^{2}—undifferentiated intrusive rocks; ^{3}—microdiorite/diorite; ^{4}—trachyte; ^{5}—quartz arenite; ^{6}—quartzite.

Table 1: *(Continued)*

Sample	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60	61	62
Quartz (monocrystalline)		.	.	.	.	.	.	.	23.3	.	.	2.0	.	-	6.7	.	20.7	.	0.6	5.7
Quartz (polycrystalline)	.	.	.	.	.	.	.	-	5.0	-	-	-	-	-	-	.	-	-	-	-
K-feldspar	.	.	.	-	-	.	.	-	tr	.	-	-	-	-	0.7	-	-	-	-	-
Plagioclase	.	.	.	.	.	.	.	.	37.0	.	.	12.7	-	-	9.7	-	42.0	.	36.3	55.7
Lithic fragments																				
Andesite	-	-	.	-	-	-	-	-	1.0	.	-	3.3	.	.	12.0	.	7.0	.	44.3	29.0
Rhyolite–dacite	.	.	.	.	.	.	.	.	29.7	.	.	23.1	.	.	23.4	.	20.7	.	10.3	3.3
Shards–pumice	-	.	-	-	-	-	-	-	-	-	-	tr	-	.	-	-	-	-	6.0	0.6
Tuff–tuffite	.	.	.	-	-	-	-	-	-	-	.	49.0	.	.	38.1	.	7.7	-	-	-
Schist	-	.	-	-	-	-	-	-	1.0	-	-	4.3	-	.	-	-	-	.	-	
Other lithic fragments	-	.	^{4}	-	^{2}	^{3,1}	-	-	3.0	-	^{1,5,6}	4.7	^{5,3}	^{3,5,6}	9.4	^{1,3}	2.0	-	2.0*	5.7*
Accessory minerals																				
Zircon	.	.	.	-	.	.	-	.	tr	.	.	-	.	-	-	.	-	.	-	-
Sphene	-	.	-	-	-	-	-	-	tr	.	-	-	-	-	-	-	tr	-	tr	-
Garnet	-	.	-	.	.	-	-	.	tr	-	-	tr	.	-	tr	-	tr	.	-	tr
Epidote	-	.	-	-	-	-	-	-	tr	-	-	-	-	-	-	-	tr	-	tr	tr
Allanite	-	.	.	-	.	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Tourmaline	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Clinopyroxene	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	tr	-
Orthopyroxene	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Muscovite	.	-	-	-	-	-	-	-	-	.	.	-	.	-	-	.	-	-	-	-
Biotite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	.	-	.	-	-
Chlorite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Ore	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Apatite	.	.	-	-	.	-	-	.	-	-	-	-	-	-	-	-	-	-	-	-
Matrix of the whole rock	X	X	X	X	X	X	X	X	29.2	X	X	X	X	X	X	X	34.8	X	X	X
Approximate grain size	cg	cg	cg	mg	fg	cgl	mg	fg	cg	cg	cgl	cgl	cgl sl	cgl	cgl	cg	cgl	cg	cgl	mg

. present; - absent; * – undifferentiated volcanic fragments; ? – identification uncertain due to recrystallization and deformation; X – not established; tr – trace; fg – fine-grained sandstone(ss.); mg – medium-grained ss.; cg – coarse-grained ss.; cgl – conglomerate; ^{1} – granite; ^{2} – undifferentiated intrusive rocks; ^{3} – microdiorite/diorite; ^{4} – trachyte; ^{5} – quartz arenite; ^{6} – quartzite.

Table 1: *(Continued)*

Sample	63	64	65	66	67	68	69	70	71	72	73	74	75	76	77	78	79	80	81
Quartz (monocrystalline)	0.3	.	.	.	-	.	.	17.3	1.5	2.3	4.3	.	3.5	.	.	.	9.3	8.0	0.7
Quartz (polycrystalline)	-	-	-	.	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
K-feldspar	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	.	-	-	-
Plagioclase	34.0	.	.	.	.	.	-	41.0	37.0	35.0	24.3	.	39.0	.	.	-	41.3	54.0	52.0
Lithic fragments																			
Andesite	20.3	.	.	.	.	.	.	14.0	-	23.0	27.0	.	-	.	.	.	23.0	16.3	30.7
Rhyolite–dacite	8.7	-	.	.	.	.	.	26.7	1.0	9.0	9.0	.	3.0	.	.	.	13.3	15.3	10.7
Shards–pumice	2.3	-	-	-	-	.	-	-	60.5	13.0	7.7	-	53.5	.	-	-	7.7	tr	-
Tuff–tuffite	14.0	-	-	-	-	.	-	-	-	13.3	22.0	-	-	-	.	.	2.0	-	4.0
Schist	-	-	-	-	-	.	-	-	-	-	-	-	-	-	-	.	-	-	-
Other lithic fragments	11.3*	^{5}	-	-	-	^{3}	^{3}	1.0*	-	4.3*	5.7*	-	-	^{6}	-	-	3.3*	6.3*	2.0*
Accessory minerals																			
Zircon	-	.	-	-	-	.	.	-	-	-	tr	.	-	-	.	-	-	-	-
Sphene	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Garnet	tr	-	-	.	-	-	.	-	-	-	tr	.	-	-	.	.	-	-	-
Epidote	-	-	-	-	-	-	-	tr	-	-	tr	-	-	-	-	-	-	-	-
Allanite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Tourmaline	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Clinopyroxene	-	-	-	-	-	-	-	tr	-	tr	-	-	-	-	-	-	-	tr	-
Orthopyroxene	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Muscovite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Biotite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Chlorite	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	tr	-
Ore	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Apatite	-	-	.	-	-	.	-	-	1.0	-	-	.	-	-	-	-	.	-	-
Matrix of the whole rock	X	X	X	X	X	X	X	X	86.0	X	80.4	X	89.6	X	X	X	X	X	X
Approximate grain size	cgl	mg	cgl	fg	cgl	cgl	cgl	cgl	tuff mg	cgl	cgl	cg	tuff mg	mg	mg	cg	cg	cg	cgl

. present; - absent; *—undifferentiated volcanic fragments; ?—identification uncertain due to recrystallization and deformation; X—not established; tr—trace; fg—fine-grained sandstone(ss.); mg—medium-grained ss.; cg—coarse-grained ss.; cgl—conglomerate; ^{1}—granite; ^{2}—undifferentiated intrusive rocks; ^{3}—microdiorite/diorite; ^{4}—trachyte; ^{5}—quartz arenite; ^{6}—quartzite.

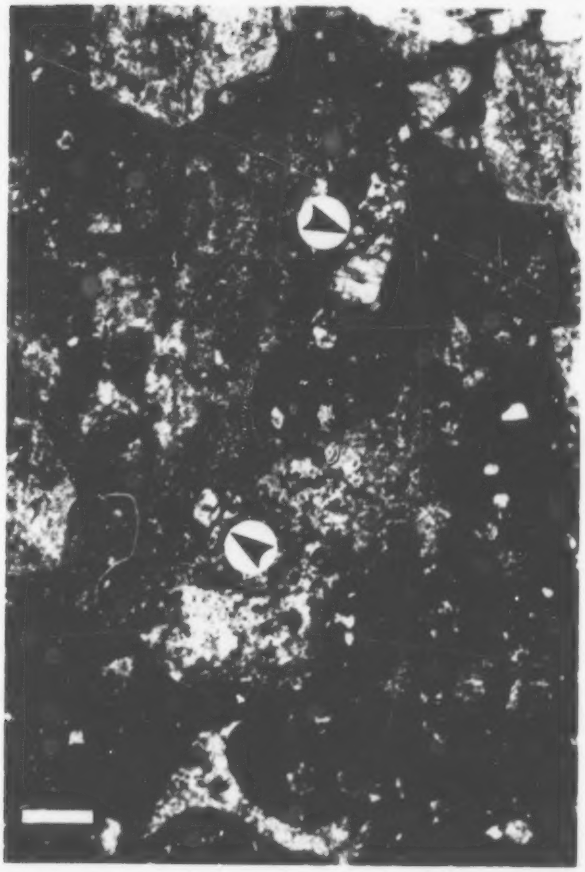

Plate 6. *Andesitic fragment containing augite phenocrysts (arrowed); note T-shaped felsic shard; plane-polarized light photomicrograph of sample 61; scale bar = 0.25 mm.*

Plate 7. *Vesicular pyroclast pebble containing skeletal plagioclase crystals; the vesicles are filled with chlorite; plane-polarized light photomicrograph of sample 72; scale bar = 0.05 mm.*

present in anomalous arkosic sandstone samples 7 and 10 (Figure 2), which consist dominantly of plagioclase feldspar and the occasional garnet inclusions; clasts of medium- to fine-grained granite are uncommon.

Sedimentary rocks. Angular fragments of intrabasinal tuff, tuffaceous sandstone and siltstone, occur in variable (usually trace) amounts in the deposits of the Connecting Point Group. In samples 54, 57, 63, 72 and 73 they are particularly abundant (Table 1; see also Plates 12 and 13 in Knight and O'Brien, 1988). There is also a minor proportion of well-rounded tuff and tuffaceous fragments, which may represent an extrabasinal component. Rare extrabasinal fragments of quartz arenite occur locally (Plates 12 and 13).

Detrital mineral grains

Quartz. Detrital quartz is present as mono- and polycrystalline grains, ranging from angular and subangular to less commonly well rounded. Polycrystalline grains usually have polygonal intercrystalline boundaries. Rounded monocrystalline quartz is often embayed, suggesting that the roundness was caused by magmatic corrosion rather than being related to sediment transport. The quartz component in the sediments, clearly decreases upward through the succession (Figure 3).

Feldspar. Plagioclase in angular to subrounded grains is the dominant feldspar. Corrosion-rounded plagioclase phenocrysts in accompanying volcanic clasts suggest that the rounded plagioclase grains may have inherited their shape. Plagioclase grains are commonly replaced by albite and K-feldspar or altered to white mica, chlorite, calcite, sphene and epidote. Orthoclase (microperthites) and microcline occur locally.

Plagioclase-rich arkoses and feldspar, ranging from 82 to 93 percent of the sand fraction, are represented by two samples (samples 7 and 10) in the Love Cove Group and lowest unit (C-1; Figure 2) of the Connecting Point Group. Both occur in facies H sandstones of Knight and O'Brien (1988).

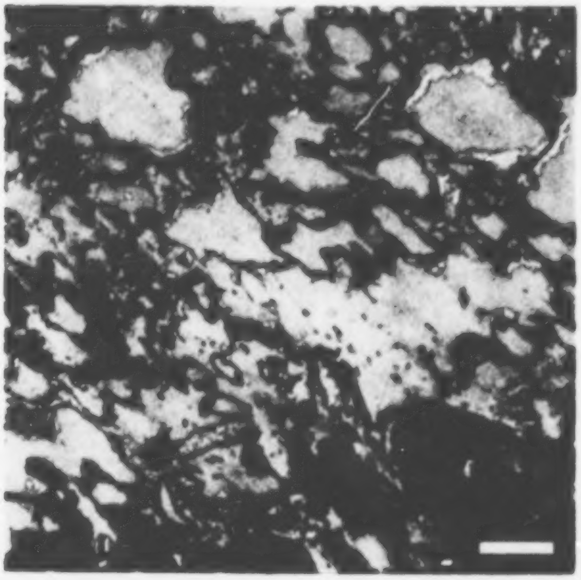

Plate 8. *Pumice fragment of andesitic-basaltic composition. Glass has mainly been replaced by chlorite. Secondary quartz, epidote and sphene are also present. The fine-grained, granular sphene outlines the margins of the vesicles. Bedding is inclined at an angle of around 30° to the right; plane-polarized light photomicrograph of sample 2: scale bar = 0.05 mm.*

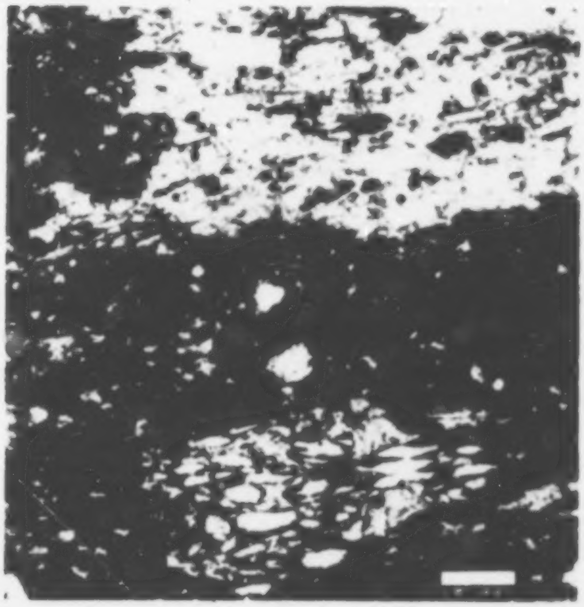

Plate 9. *Jagged pumice fragments showing bedding-parallel flattening; because some of the pumice particles experienced very little deformation, it is inferred that tectonic origin of this texture is unlikely; plane-polarized light photomicrograph of sample 2; scale bar = 0.25 mm.*

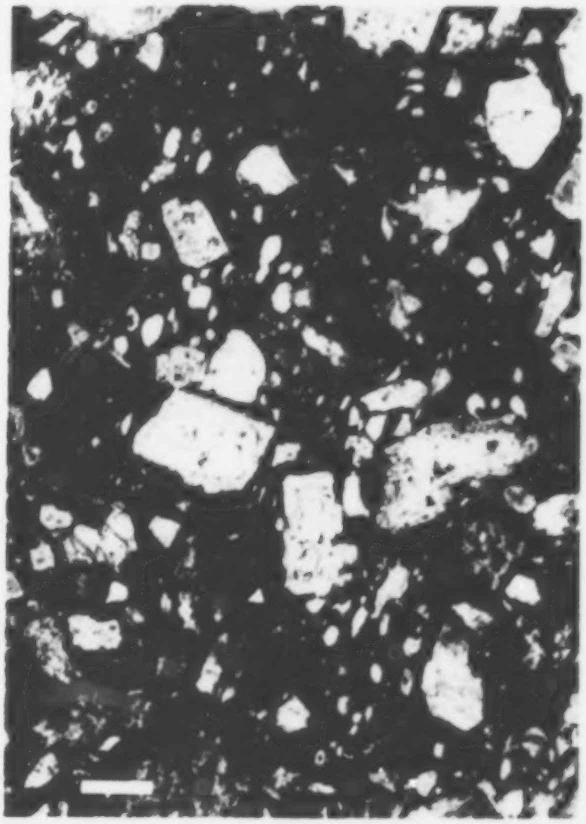

Plate 10. *Multi-lobate, andesitic lapilli; the vesicles are mostly filled with chlorite and plagioclase phenocrysts are overgrown by albite(?); plane-polarized light photomicrograph of sample 61; scale bar = 0.25 mm.*

Accessory mineral grains. Garnet is the most common accessory mineral and occurs as euhedral, round and angular grains, ranging in size from 0.01 to 0.12 mm (Plate 14). Garnet has been found associated with muscovite–biotite schist fragments. Small garnet inclusions occasionally also occur in plagioclase crystals associated with the intermediate volcanic and intrusive lithic fragments. Zircon, apatite and allanite are less common; their average size is 0.05, 0.08 and 0.16 mm respectively. Sphene, epidote, opaque minerals (principally pyrite), and magnetite are also present, but it is very often impossible to draw a distinction between the detrital grains and the secondary crystals. Augite and tourmaline are the least common accessory minerals.

Matrix

The dominant component of most of the sediments in the Connecting Point Group is a ubiquitously altered, fine-grained matrix (<0.0626 mm). It comprises up to 86 percent of some samples (Table 1). Cryptocrystalline quartz, feldspar, sericite, muscovite, chlorite and calcite, commonly replace

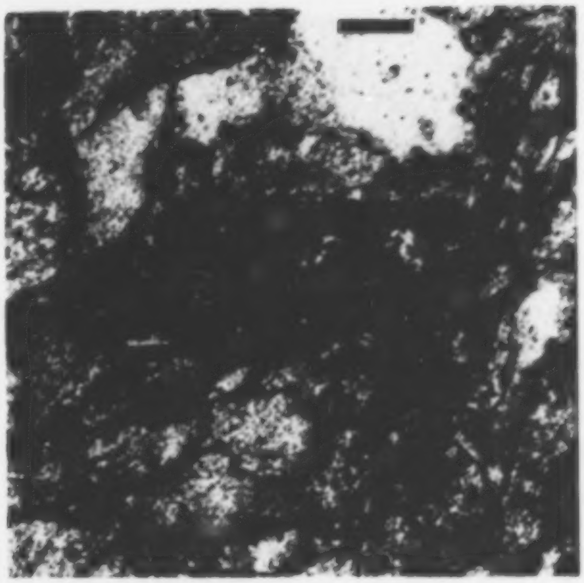

Plate 11. *Slightly deformed fragment of biotite–muscovite schist enclosed in devitrified tuffaceous matrix; plane-polarized light photomicrograph of sample 17; scale bar = 0.25 mm.*

the original matrix. Epidote, clinozoisite, sphene, prehnite and pyroxene are other, less common, secondary minerals. Relicts of shards and pumice occur locally and suggest that the matrix in the sandstones and conglomerates was originally felsic lithic and crystal tuff material.

Relationship Between Existing Lithostratigraphy and Petrography

The petrographic study generally confirms the lithostratigraphic correlations between sections of the Connecting Point Group, originally established by Knight and O'Brien (1988). It also identifies a compositional change of lithic clasts in the sandstones, from dominantly felsic below the mixtite, to andesite-dominant, at and for 1500 m above the mixtite (Figure 3).

The granule conglomerate from the Willis Reach section (sample 59, Figure 2) has a different composition from the other samples of the granule conglomerate (Table I), collected from the mixtite horizon, with which it is correlated. This disparity does not necessarily imply an incorrect correlation. It may merely reflect a compositionally heterogeneous nature of the mixtite sheet, which in turn reflects the lithological inhomogeneities of the source sediments, which were present along the slope of the Eastport basin, prior to the remobilisation and emplacement of the mixtite.

Plate 12. *Sand-size grains of microdiorite (above) and quartz arenite; cross-polarized light photomicrograph of sample 56; scale bar = 0.25 mm.*

Pyroclastic versus Epiclastic Origin

Petrographic textures and sedimentary structures indicate that most of the sediments in the Love Cove and Connecting Point groups were derived by reworking of dominantly pyroclastic deposits of a penecontemporaneous volcanic arc apron. Autochthonous tuffs in the Connecting Point Group (samples 71 and 75, Figure 2) show no petrographic evidence of welding. However, some of the deposits in the upper Love Cove Group and the lowest Connecting Point Group may have formed from pyroclastic surges (cf. Ayres, 1982). The existence of possible eutaxitic textures provide inconclusive evidence of welding in thin, fine-grained, glassy beds interpreted as tuffs by Knight and O'Brien (1988, facies B) (Plate 9).

Plate 13. *Sand grain of quartz arenite; cross-polarized light photomicrograph of sample 56; scale bar = 0.05 mm.*

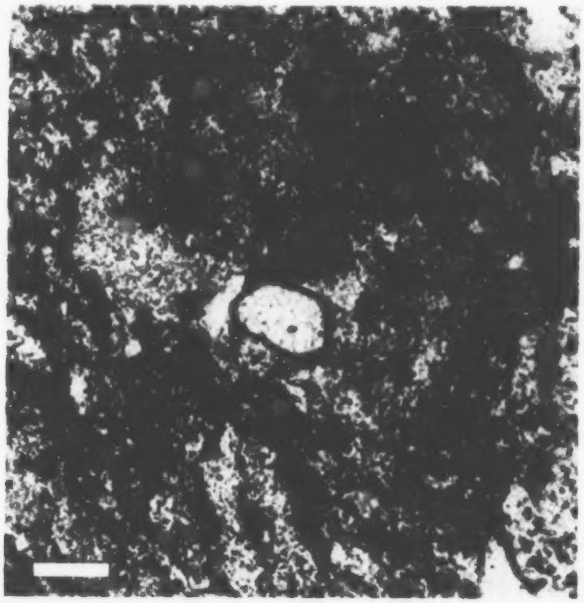

Plate 14. *Euhedral grain of garnet set in devitrified tuffaceous matrix; garnet also appears as a constituent of biotite–muscovite schist fragments; plane-polarized light photomicrograph of sample 17; scale bar = 0.25 mm.*

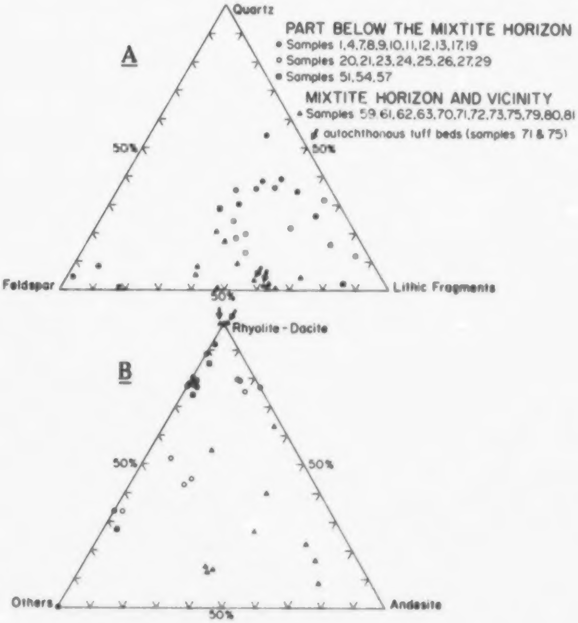

Figure 3. *(A) Standard trigonal diagram summarizing petrographic composition of the detrital fraction in the upper part of the Love Cove Group and the overlying Connecting Point Group. The diagram illustrates an upward decrease of the quartz component ("immaturing-upwards"); (B): discrimination of lithic fragments in the same samples as in A shows an increase in proportion of andesitic detritus at the mixtite horizon.*

Provenance of the Sediments of the Love Cove and Connecting Point Groups

Felsic pyroclastic material, represented by a fine-grained matrix and rhyolitic–dacitic lithoclasts, predominate the entire upper part of the Love Cove Group, and the overlying Connecting Point Group. It records a period of almost continuous felsic pyroclastic volcanism, which released vast volumes of ash into the adjacent Eastport basin. Although a few autochthonous crystal tuff beds have been examined from both below and above the mixtite horizon, (samples 71 and 75, Figure 2), most of the fallout tephra underwent remobilisation prior to final deposition on the submarine apron (Knight and O'Brien, 1988). The resultant tuffaceous deposits are heterolithic and are enriched in the lithic volcanogenic material, as well as in quartz and plagioclase of both pyroclastic and epiclastic origin. This aspect of resedimentation is best demonstrated by the mixtite, which comprises large blocks of tuffite and volcanigenic conglomerate (Plates 12 and 13 in Knight and O'Brien, 1988), and where the matrix contains abundant vitric, felsic fragments (Table 1). The autochthonous tuff bed (sample 71, Figure 2), represents a possible source of the felsic vitric material, which is present in the mixtite directly above.

The upward decrease in the proportion of quartz may reflect a significant decline in, 1) the proportion of epiclastic detritus in the source area, and 2) the proportion of material derived from erosion of basement rocks. This 'immaturing-upwards' may also reflect a decrease in the proportion of sediment that underwent textural modification in fluvial, deltaic or shallow marine settings prior to the resedimentation. This consequently implies an upward increase in the amount of pyroclastic detritus in the turbidites, and is supported by the conspicuously abundant tuffaceous material associated with the mixtite horizon, where it appears as both autochthonous tuff beds (samples 71 and 75, Figure 2), clasts (samples 69 and 70, Figure 2), and as matrix in the conglomerates (samples 72 and 73, Figure 2). This overall trend probably, partly reflects the generation of large quantities of pyroclastic material in the vicinity of the arc, overwhelming the basin processes and resulting in slope instabilities, which readily led to downslope remobilisation. In addition, it also may reflect the widening of the arc basin with time (Knight and O'Brien, 1988), explaining the upward change from a low-efficiency deep-water fan setting, to fans composed of well-developed channel and levée complexes. When the basin was narrow with steep sides, sediment reworked by nearshore and fluvial processes along the basin margins was readily moved basinward, down the slope. This scenerio is supported by mega-turbidites of plagioarkosic sandstones (see next section) that occur in this succession. As a result when the basin widened and slopes declined, these zones of shallow water sediment reworking would be effectively isolated by the apron from the basin depocenter, so that progressively, less reworked detritus was remobilised and transported basinward.

Andesitic and associated dioritic detritus is most conspicuous in the mixtite horizon and the deposits above it. Andesite fragments range from vitric to lithic, and their variable roundness may indicate that they are a mixture of epiclastic and pyroclastic material. This sudden increase in the proportion of the andesitic detritus suggests a change in the composition of volcanism in the vicinity of the Eastport basin. It appears to have coincided with an episode of basin extension and accompanying mafic intrusions (Knight and O'Brien, 1988). The andesitic material is accompanied by pyro- and epiclastic rhyolite–dacite detritus showing that felsic volcanism, although less important, was almost continuously active during the sedimentation of the Connecting Point Group.

As is the case of most resedimented volcaniclastic rocks, (e.g., Schmincke and von Rad, 1979; Fisher and Schmincke, 1984; Cas and Wright, 1988) the Connecting Point Group contains various amounts of non-volcanic components. Fragments of intrusive, metamorphic and clastic sedimentary rocks were incorporated during the formation of the tuffaceous turbidites, the debris flow conglomerates and the olistostrome deposits. Generally, the metamorphic detritus appears in trace amounts; however, in the tuffite (sample 17, Figure 2), sand-size schist fragments are exceptionally abundant (8.5 percent). The enrichment might have occurred through winnowing of the vitric material and/or by incorporation of new detritus.

The origin of these metamorphic, dioritic, granitic and quartz arenite fragments is uncertain. They are either, 1) accidental constituents captured from an eruptive vent during an explosive eruption (Fisher and Schmincke, 1984; Suzuki-Kamata, 1988), or 2) the product of erosion of exposed areas of continental crust and/or island-arc crust.

Sedimentary rock fragments are predominantly represented by intrabasinal tuff and tuffaceous rock fragments (see also Plates 12 and 13 in Knight and O'Brien, 1988).

Compositionally Distinct Sandstones

Petrographically and texturally distinct arkosic sandstones occur in two units of graded, planar-stratified sandstones (facies H, Knight and O'Brien, 1988), in the first 800 m of the section on Long Reach Island (samples 7 and 10, Figure 2). Throughout the section other units of this facies have a mineralogy similar to the rest of the sequence. The arkoses are well sorted with angular to subrounded grains and contain lithic grains of microdiorite. The facies units have restricted lateral distribution in the fine grained, basinal and distal fan deposits. Locally, they fill channels within levée deposits. The two examples described, occur within distal fan turbidites of the lower part of the succession that are compositionally lithic arenites and have a predominantly felsic component. This suggests that the arkose were derived from a distinct source, probably a pluton, and when transported as mega-turbidites, they were able to bypass the volcaniclastic apron, and not loose their mineralogical identity by dispersal in shallow- or deep-water settings. The source may have been a subvolcanic dioritic–microdioritic pluton, which was exposed and eroded by a single river that constructed a delta or shoreline sand body of distinct detrital composition. The sands were remobilised soon after and travelled as mega-turbidites to basinal depths.

Subaerial or Submarine Eruptions

The widespread presence of the felsic and intermediate epiclastic detritus in the succession of the Eastport basin and the local occurrence of reworked arkosic detritus (samples 7 and 10; Figure 2) indicate a subaerially exposed volcanic-intrusive edifice, as the source of a substantial proportion of the detritus released into the basin. The widely distributed but uncommon schist, granite and quartz arenite detritus additionally suggest an emergent terrane having a possible metamorphic basement to the volcanic-subvolcanic edifice.

Extensive pyroclastic volcanism is not likely to begin until a volcano is either above or close to sea level (Ayres, 1982). In particular, the subaerial parts of volcanoes of silicic to intermediate composition, are known to yield a great amount of the pyroclastic detritus (Fisher and Schmincke, 1984; Cas and Wright, 1988). An abundant supply of unconsolidated pyroclastic sediments is critical in explaining the abundance of the resedimented facies in the Eastport basin.

The vitric material recognized in the Connecting Point Group is characterised by a high degree of vesiculation and the andesitic lapilli commonly contain quench plagioclase, indicative of rapid cooling (Gelinas and Brooks, 1974). These features additionally support the subaerial and/or shallow-water nature of the felsic to intermediate volcanic eruptions (Heiken, 1972; Fisher and Schmincke, 1984). Hyaloclastic material, which is diagnostic of submarine eruptions (Fisher and Schmincke, 1984), has not been identified in the group.

Comments on Sedimentary Processes

A significant proportion of the detritus in the Connecting Point Group, probably underwent recycling in fluvial or shallow-marine settings prior to transportation onto the volcaniclastic apron. This is suggested by numerous well rounded clasts of various lithologies (Plate 2) and locally by good sorting of the sand grains (sample 10; Figure 2). However, it is noteworthy that a significant degree of roundness can also be achieved during pyroclastic eruption because of the high concentration of pyroclasts passing through the eruptive vent (Cas and Wright, 1988).

In volcaniclastic sediments, the rate of glass divitrification is sufficiently rapid to produce a significant amount of smectite (Vessel and Davis, 1981; Fisher and Schmincke, 1984). It is inferred that the presence of this expandable clay in the sediments of the Eastport basin might have considerably increased the cohesion factor of the gravity flows during resedimentation and hence facilitated development of high-density turbidity currents. In addition, smectite-rich sediments, as well as ash-water mixtures, have thixotropic properties and, therefore, slopes composed of such sediments would be highly unstable and prone to collapse (e.g., Busby-Spera, 1988).

CONCLUSIONS

The results of this petrographic study strongly support the volcanic arc-related geotectonic setting of the Eastport basin and the sedimentation of the Connecting Point Group, proposed by Knight and O'Brien (1988). Penecontemporaneous volcanic rocks ranging from rhyolites to andesites suggest a mature island-arc complex with a substrate of continental and/or island-arc crust, including medium- to high-grade metamorphic rocks (cf. Hamilton, 1988) that supplied subordinate metamorphic, plutonic and quartz arenite detritus.

More field work is necessary in order to carry out clast counts in the mixtite horizon and the adjacent coarse-grained deposits. This would give a more precise estimate of the proportions of various lithic components present in the Connecting Point Group. Further detailed petrographic work and bulk geochemical analyses on the clasts will enable a better understanding of the volcanic source area and its comparison with volcanic rocks of the contemporaneous Love Cove Group. Collection of additional samples of sandstones from selected horizons will supplement and confirm the existing petrographic data of the Connecting Point Group. Microprobe analysis of detrital garnet, feldspars and pyroxene should help to better define composition and origin of source rocks in the volcanic arc and in its substrate.

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ASPECTS OF THE MID-PALEOZOIC MAGMATIC HISTORY OF THE SOUTH-CENTRAL HERMITAGE FLEXURE AREA, NEWFOUNDLAND

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ABSTRACT

The Burgeo intrusive suite, a mainly synkinematic polyphase complex containing dated Silurian plutons, intrudes lower to mid-Paleozoic rocks on the southwest coast of Newfoundland. It comprises gabbro, quartz diorite, biotite ± hornblende tonalite, granodiorite and granite, and biotite ± muscovite granite. At its western and eastern limits, respectively, the suite is intruded posttectonically by the Devonian Chetwynd Granite and the possibly synchronous François granite. These are high-level granites that cross-cut adjacent mylonite zones developed within the Burgeo intrusive suite and its country rocks.

The Burgeo intrusive suite shows a continuum in geochemical variation from 49 percent to 79 percent SiO_2*, similar to that seen in calc-alkaline suites, but there is a greater abundance of high-silica and alkali-enriched granites. It is probable that the mafic rocks and lower-silica granitoids are related to the subduction zone and lower-crust magmatism and subsequent fractionation, whereas the higher-silica granites reflect high-level crustal melts. The François and Chetwynd granites have restricted* SiO_2 *compositions (mainly >71 percent). Both granites are strongly depleted in Sr and ferromagnesian elements and the François granite is enriched in Rb, U and Th. The granites are probably high-level crustal melts, which have fractionated biotite, plagioclase and K-feldspar. Small intrusions of similar lithology are amongst the latest phases of the Burgeo intrusive suite.*

Known mineralization associated with the Burgeo intrusive suite is concentrated in the Grey River area, where extensive wolframite-bearing quartz veins are present. A small gold occurrence has also been reported from this area. Molybdenite mineralization is locally developed in aplites of the suite. The most differentiated parts of the posttectonic intrusions are potential targets for granophile element mineralization.

INTRODUCTION

Voluminous granitoid batholiths are the dominant component of the Appalachian orogen in south-central Newfoundland (Figure 1). These composite intrusive suites were emplaced into mainly lower to mid-Paleozoic greenschist and amphibolite facies country rocks, during the polyphase tectonothermal event that affected the region in the Silurian (e.g., O'Brien *et al.*, 1986; Dunning *et al.*, 1988). The country rocks, which are almost completely enveloped by intrusions, are mostly volcanic and sedimentary protoliths. They include elements of the Dunnage and Gander zones of the Appalachian mobile belt (Williams, 1979; Williams *et al.*, 1988; Williams *et al.*, 1989) and the Hadrynian to Cambrian (Dunning and O'Brien, *in press)* Grey River enclave (Blackwood, 1985a), currently of uncertain affinity. In southern Newfoundland, these rocks define the sinuous oroclinal structure known as the Hermitage Flexure (Williams *et al.*, 1970). In addition to composite batholiths, smaller plutons of epizonal, high-silica granite, are widely spaced along the entire southern coast (Dickson *et al.*, 1988) and cross-cut regional fabrics in both the country rocks and the earlier granites, and are of Devonian or presumed Devonian age.

The following discussion deals with the geology and geochemistry of three mid-Paleozoic intrusions from the south-central part of the Hermitage Flexure area. These are: the synkinematic, polyphase Burgeo intrusive suite[1], the posttectonic Devonian Chetwynd Granite (Cooper, 1954) and the posttectonic François granite (Poole *et al.*, 1985). These three intrusions have been recently mapped at 1;50,000 scale (see O'Brien *et al.*, 1986) and systematic geochemical sampling on a 4-km^2 grid is essentially complete, providing a total of 822 samples, which have been analyzed for major- and trace-elements. U–Pb age determinations on two phases of the Burgeo intrusive suite have been completed (Dunning *et al.*, 1988). Geochronological and geochemical studies and detailed mapping are continuing.

GEOLOGY OF THE INTRUSIONS

Burgeo Intrusive Suite

The Burgeo intrusive suite (Figure 1) is a polyphase intrusive body of batholithic proportions (circa. 3000 km^2; O'Brien and Dickson, 1986) composed of a continuum of compositions, from gabbro through tonalite and granodiorite

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This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

[1] "Burgeo intrusive suite" is a new name to be proposed elsewhere. It replaces the Burgeo Batholith of Williams (1978) and the Burgeo granite of O'Brien and Tomlin (1984, 1985) and Dickson *et al.* (1985).

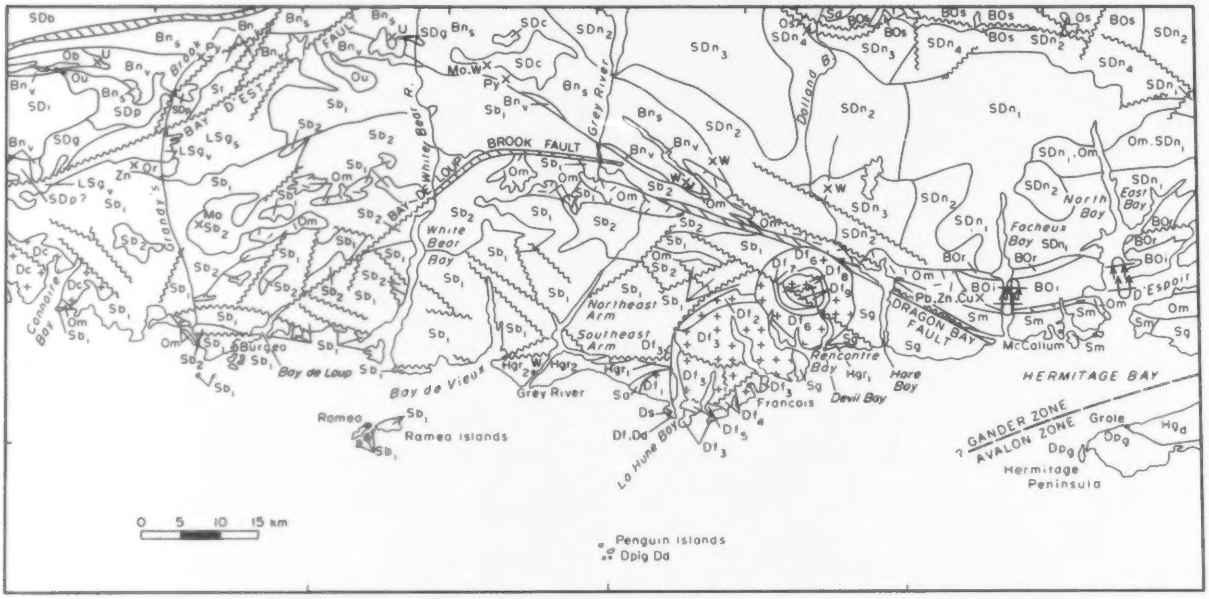
BROOK FAULT
BAY D'EST FAULT
BAY DE LOUP
DRAGON BAY FAULT
Grey River
White Bear R.
Dolland
White Bear Bay
Northeast Arm
Southeast Arm
Facheux Bay
North Bay
East Bay
D'Espoir
Rencontre Bay
Hare Bay
Devil Bay
Francois
La Hune Bay
Burgeo
Bay de Loup
Bay de Vieux
Romeo
Romeo Islands
Connoire Bay
McCallum
HERMITAGE BAY
Hermitage Peninsula
Grole
GANDER ZONE
AVALON ZONE
Penguin Islands
Dpig Dd
Pb,Zn,CuX
Mo,WX
0 5 10 15 km

Figure 1. *Geology of the Burgeo (11P) map area; modified from O'Brien et al. (1986).*

LEGEND

DEVONIAN

Ds	*Red and green conglomerate, sandstone, shale and limestone*
Dd	*Undivided early and late, massive diabase dykes*
Df	FRANCOIS GRANITE: *pink, massive, biotite granite;* Df_1*, buff, medium grained, feldspar porphyritic granite;* Df_2*, coarse grained equigranular granite;* Df_3, Df_6*, coarse grained, K-feldspar porphyritic granite;* Df_4, Df_8*, buff, medium grained, feldspar porphyritic adamellite and granite;* Df_5*, pink to white, fine grained, quartz-feldspar-biotite porphyritic granite;* Df_7*, medium grained, K-feldspar porphyritic granite;* Df_9 *medium grained equigranular granite*
Dc	CHETWYND GRANITE: *medium to coarse grained, equigranular to K-feldspar porphyritic, biotite granite*
Dpg	PASS ISLAND GRANITE: *pink, medium to coarse grained, K-feldspar porphyritic, biotite-hornblende granite*
Dpig	PENGUIN ISLANDS GRANITE: *pink, medium grained, K-feldspar porphyritic, biotite granite*
Dg	*Undivided, medium to fine grained, biotite ± hornblende granite*

SILURIAN–DEVONIAN

SDg	*Medium grained muscovite granite and felsite*
SDI	IRONBOUND MONZONITE: *massive to locally foliated, medium grained, plagioclase porphyritic, biotite ± hornblende monzonite with minor granodiorite and granite*
SDp	PETER SNOUT GRANITE: *massive to weakly foliated, fine to medium grained, equigranular, biotite ± muscovite granite*
SDc	COCHRANE POND GRANITE: *massive to weakly foliated, fine to medium grained, muscovite ± biotite granite*
SDn	SDn_1*, weakly foliated, medium grained, equigranular to porphyritic, biotite ± muscovite granodiorite, with minor garnet-muscovite granite;* SDn_2*, massive to weakly foliated, medium to coarse-medium grained, K feldspar porphyritic, biotite granite;* SDn_3*, massive to weakly foliated, medium to coarse-medium grained, K-feldspar porphyritic, biotite-muscovite granite;* SDn_4*, strongly deformed to mylonitic, medium to coarse grained, K-feldspar porphyritic, biotite granodiorite, and medium grained equigranular muscovite-garnet granite*

SILURIAN

Sd	Dolland Pond Formation: *cleaved pebble conglomerate and breccia, interbedded with shale and sandstone*
Sa	*Fine to coarse grained, hornblende gabbro, including hornblendite, diorite and amphibolite*
Sb	BURGEO GRANITE: Sb_1, *coarse-medium to coarse grained, massive to foliated, feldspar porphyritic, biotite ± hornblende granodiorite, tonalite and gabbro;* Sb_2, *medium grained, foliated equigranular to porphyritic, biotite ± muscovite ± garnet granodiorite and granite, locally containing abundant migmatite inclusions*
Sm	McCALLUM GRANITE: *foliated to mylonitic, medium to coarse grained, equigranular to feldspar porphyritic, biotite granite and granodiorite*
Sg	GAULTOIS GRANITE: *mainly coarse grained, strongly foliated to mylonitic, feldspar porphyritic, biotite granodiorite and granite, including minor, medium grained, equigranular granite*
St	TOP POND TONALITE: *foliated, medium grained, equigranular, hornblende ± biotite tonalite*

LA POILE GROUP

LSg	Georges Brook Formation: LOg_V, *rhyolite flows, ash-flow tuffs, agglomerate, reworked volcaniclastics, minor mafic flows, pyroclastics and dikes;* LOg_S, *fine to coarse grained, locally cross-stratified sandstone, including quartz-pebble and quartz-cobble conglomerate*

ORDOVICIAN?

Om	*Migmatite, agmatite, paragneiss and associated granitoids; locally includes migmatized metasedimentary rocks of known Ordovician age*
Ob	BAGGS HILL GRANITE: *massive to weakly foliated, quartz-feldspar porphyritic granite, granodiorite and quartz-feldspar porphyry*

ORDOVICIAN

Os	Spruce Brook Formation: *strongly foliated psammitic and semipelitic schist and minor schistose felsic volcanic rocks; locally extensive, cleaved, quartz-pebble conglomerate and quartzite*

BAY DU NORD GROUP

Bn	Bn_V, *felsic pyroclastic rocks, rhyolite, and welded tuff associated with mafic volcanic rocks and clastic sedimentary rocks;* Bn_S, *psammite, semipelite and pelite, including quartz-rich sandstone and quartz-pebble conglomerate*

BAIE D'ESPOIR GROUP

BOr	Riches Island Formation: *semipelitic schist, including thin to thick bedded psammitic schist and minor quartzite*
BOi	Isle Galet Formation: *felsic volcanic rocks and overlying graphitic pelite, arenite and minor tuff, conglomerate, limestone and ampbibolite*
BOs	Salmon River Dam Formation: *psammite and pelite containing minor calcareous psammite and limestone*

ORDOVICIAN OR EARLIER

Or	ROTI GRANITE: *foliated, equigranular, medium grained, biotite granodiorite and quartz-feldspar porphyry.*
Ou	*Layered metagabbro and metadiabase; minor pyroxenite and peridotite; rare mafic volcanic rocks*

HADRYNIAN TO CAMBRIAN

Hgr	GREY RIVER ENCLAVE: Ogr_1, *felsic tuff and minor metasedimentary rocks;* Ogr_2, *migmatized pelitic and psammitic schist, mafic schist and amphibolite; minor gabbro, peridotite and granitoid gneiss*

HADRYNIAN

Hgd	GROLE DIORITE: *medium to coarse grained, hornblende-pyroxene gabbro and hornblende-biotite diorite, and quartz diorite*

to granite. The oldest rocks, which are intruded and variably assimilated by several younger plutons, are the most mafic in composition and include mainly hornblende gabbro, biotite–hornblende tonalite and quartz diorite; these underlie small areas and are unseparated from other phases of Sb_1, on the regional map (Figure 1). The gabbroic rocks are most common on the Ramea Islands and the Burgeo Islands, where they are associated with diorite. Diabase dykes are common on Ramea Great Island, where brecciated and mylonitic gabbro and granite are cut by massive diabase dykes. Away from its southernmost exposures, diabase dykes in the Burgeo intrusive suite are comparatively rare. The concentration of possibly several generations of dykes, along the coast, may reflect a fundamental weakness in the crust in this area. Foliated quartz diorite and granodiorite form a separate pluton in the northern part of the suite (part of Unit Sb_1), west of White Bear River (see O'Brien and Dickson, 1986). Its genetic relation to other parts of the suite is enigmatic.

The most diagnostic and widespread lithology within the Burgeo intrusive suite, which comprises most of Unit Sb_1, is coarse grained, K-feldspar-porphyritic, biotite ± hornblende granite and granodiorite. In the west, these rocks are divisible into K-feldspar and plagioclase phenocryst-rich and sparsely porphyritic phases, a biotite-rich (up to 30 percent modal biotite) phase, as well as a unit rich in cognate mafic xenoliths (O'Brien and Tomlin, 1985). The biotite-rich porphyritic phases are, in places, compositionally similar to the most differentiated parts of the older elements of the suite, possibly supporting a genetic relationship between them. Variably sheared, pink granite intrudes some of the above rocks and forms a separate unit in the south-central part of the suite between Bay de Loup and Burgeo.

The youngest, regionally extensive component is medium grained, biotite ± muscovite granite (Unit Sb_2) that outcrops in a 10 to 15-km-wide east-trending band, forming the core of the suite. Locally, small bodies of massive, coarse grained, leucocratic biotite granite, similar to that found in the Chetwynd and François granites, intrude the older phases.

Enclaves of orthogneiss, migmatite and mafic to ultramafic rocks are widespread within the Burgeo intrusive suite. In the northern part of the suite, amphibolitized gabbro and diabase and minor serpentinized ultramafic rock form a large roof pendant surrounded by granite (O'Brien and Tomlin, 1984). This unit may either be an ophiolitic remnant, correlative with similar Cambro–Ordovician mafic–ultramafic complexes to the northwest and west (e.g., Chorlton, 1980a), or relics of the amphibolite terrane of the southern Hermitage Flexure. Farther south, mafic migmatite and gneissic granite form xenoliths within the suite (O'Brien and Tomlin, 1985). These may be equivalent to either deeper level metamorphic and intrusive rocks of Silurian age or to the pre-Silurian elements of the Hermitage Flexure, exposed elsewhere in southern Newfoundland. Ovoid to tabulate mafic igneous xenoliths are common within felsic phases throughout the Burgeo intrusive suite. These may represent reworked early mafic phases of the suite, disruption of mafic rich bands produced by gravitational layering within the magma chamber, or globular enclaves formed by disaggregation of co-existing mafic phases (c.f., Vernon *et al.*, 1988).

Almost all the elements of the suite contain an inhomogeneously developed fabric, which is in many places penetrative. The main fabric in the older porphyritic granite phases has been correlated with the regional first (D_1) fabric of the adjacent Ordovician rocks (Chorlton, 1980a; Blackwood, 1984). Away from the margin of the suite, the relation of fabric development in subsequent phases, to that in the country rock is difficult to ascertain, given the inhomogeneous nature of the deformation, the uncertainty in correlating fabrics amongst intrusive units and the overall synkinematic nature of the suite. Major mylonite zones are developed within porphyritic granitoid phases along the southern boundary of the suite with the Grey River enclave, at Grey River (Blackwood, 1985a), and near its northern contact with the Ordovician Bay du Nord Group. The northern mylonite zone, which is the on-strike equivalent of the Dragon Bay Fault (Blackwood, 1985b), swings east–west from the contact, and demarcates the northern limit of migmatite within the intrusive suite east of the Bay de Loup Brook Fault (Figure 1; Dickson *et al.*, 1985).

The Burgeo intrusive suite intrudes metasedimentary and radiometrically-dated (468 Ma; Dunning *et al.*, 1986) metavolcanic rocks of the Bay du Nord Group along much of its northern contact. The latter is part of the Exploits Subzone of the Dunnage Zone (Williams *et al.*, 1988). Southeast of the Bay D'Est Fault, the suite intrudes sedimentary and minor associated volcanic rocks, assigned to the Silurian La Poile Group (O'Brien, 1983; Dunning *et al.*, 1988). The metavolcanic, metasedimentary and gneissic rocks of the Grey River enclave, lie to the south of the Burgeo intrusive suite and are in fault contact with it (Blackwood, 1985a). Possible equivalents of the suite are the Otter Point (O'Brien, 1987) and Gaultois (Colman-Sadd, 1974) granites. They are separated from it by the posttectonic François and Chetwynd granites that define the suite's eastern and western limits (see below).

Dunning *et al.* (1988) report U–Pb ages on two of the main plutons of the Burgeo intrusive suite. A coarse-grained, foliated, K-feldspar-porphyritic, biotite granite (Unit Sb_1), west of White Bear River, has a zircon (crystallization) age of 429 $^{+5}_{-3}$ Ma and a titanite (metamorphic) age of 417 ± 2 Ma. Lithologically similar granite, exposed at the contact with the Bay Du Nord Group in the Dolland Brook area, contains the D_1 fabric of the country rocks (Blackwood, 1984). Medium grained, weakly foliated, equigranular biotite–muscovite granite (Unit Sb_2), which cross-cuts the older dated phase, has a zircon age of 415 ± 2 Ma. These ages are interpreted to bracket the main period of Silurian synkinematic granite intrusion and the regional D_1 deformation of the country rocks (Dunning *et al.*, 1988). The contrasting compositions and 10 Ma age difference between the two dated plutons suggests that they may not be linked by petrogenetic processes.

Higgins (1980) reported Rb–Sr whole rock ages from undeformed, K-feldspar-porphyritic biotite granite from the Burgeo intrusive suite in the Grey River area. Three samples of this granite gave an isochron age of 405 ± 10 Ma, with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7092. One of the samples used gave a K/Ar (biotite) age of 462 ± 18 Ma. The old K/Ar age was interpreted by Higgins (1980) to be erroneous and due to excess argon.

Chetwynd Granite

The Chetwynd Granite is a posttectonic, Devonian intrusion (Dallmeyer, 1980; Chorlton and Dallmeyer, 1986) composed of leucocratic, biotite granite. Two main regionally mappable phases are recognizable: these are fine- to coarse-grained, equigranular granite and K-feldspar ± quartz-porphyritic granite, containing adamellite variants. Rapakivi texture is locally well preserved in phenocryst-rich cumulate rocks. The more quartz-rich, silicic granites occupy higher ground than the less siliceous, quartz-poor, K-feldspar-rich variety (see below). The former lithology locally cross-cuts the less differentiated phases. The granite intrudes the westernmost extension of the Burgeo intrusive suite, and cross-cuts foliations within it and all the adjacent units (Chorlton, 1978; O'Brien and Tomlin, 1985; O'Brien, 1987). The western contact of the granite is semicircular in outline, whereas in the east, the contact is more irregularly shaped and locally rectilinear (Figure 1). Although the granite is posttectonic with respect to the regional deformation, it is affected by late northwest-trending structures, which may have been instrumental in defining the shape of the granite in its eastern portions. The presence of miarolitic cavities, tuffisite, and granophyre within the granite are indicative of high-level emplacement. The occurrence of associated sanidine-bearing porphyry dykes that are compositionally similar to the granite, within adjacent country rocks, is a further indication of the granite's epizonal character. These and related rhyolite and quartz–sanidine-porphyry dykes occur up to 15 km from the Chetwynd Granite. Screens of deformed Burgeo intrusive suite, La Poile Group, and older metamorphic rocks are common within the Chetwynd Granite and range in size from a square metre to several square kilometres.

The Chetwynd Granite cross-cuts folds and fabrics in dated Silurian strata as well as structures and major faults in the pre-Silurian succession (O'Brien, 1987, 1988). The granite yields a cooling age of 372 ± 5 ($^{40}Ar/^{39}Ar$ biotite) and a U–Pb zircon crystallization age of 377 ± 20 Ma (Dallmeyer, 1980; Chorlton and Dallmeyer, 1986).

François Granite

The François granite is a high level, posttectonic granite intrusion, consisting of two overlapping, circular ring complexes that create a distinctive bilobate map pattern (Figure 1). The pluton is one of a number of posttectonic, highly differentiated, high-silica granite intrusions that form an east-trending, 350-km-long belt along the southcoast of Newfoundland (Dickson *et al.*, 1988). The granite outcrops at the eastern limit of the Burgeo intrusive suite and separates it from the Gaultois granite, an extensive synkinematic granitoid that is continuous eastward to the Hermitage Bay Fault (Blackwood and O'Driscoll, 1976). The earliest phases of the François granite include early, coarse grained, massive, equigranular (Unit Df_2) and K-feldspar-porphyritic–biotite granite (Units Df_3 and Df_6), which together form the outer rings of both lobes. A prominent buff to grey, massive, relatively biotite-rich, plagioclase–K-feldspar-porphyritic biotite granodiorite (Unit Df_8) forms a ring intrusion, adjacent to the highly differentiated aplitic core in the northeastern complex. It is the penultimate phase of intrusion in the northeastern complex and is also the granite's most mafic phase. The same rock type (Unit Df_4) forms a sill-like intrusion in the southwestern lobe. Massive, aplitic, biotite granite (Unit Df_9) forms the central portion of the northeast lobe.

The François granite and associated dykes posttectonically intrude the Bay du Nord Group, the Burgeo intrusive suite and the Gaultois granite, and host xenoliths of these units. The granite post-dates movement on both the Dragon Bay Fault zone and the mylonite zone that form the northern boundary of the Grey River enclave. Rhyolitic, aplitic, quartz–sanidine porphyry and granite dykes occur up to 10 km north of the François granite and have intruded the nearby Unit SDn_2 of the North Bay granite suite, which has been dated at 396 Ma (U–Pb zircon; Dunning *et al.* 1988). At La Hune Bay, a sequence of gently dipping, coarse grained, subaerial clastic sediments are interpreted to nonconformably overlie the François granite (Blackwood, 1985a). The sequence is assumed to be Devonian or younger in age (Blackwood, 1985a), mainly because of lithological similarity to fossiliferous Devonian strata farther east in Fortune Bay. Both the granite and its cover are cross-cut by fine grained diabase dykes.

GENERAL GEOCHEMICAL CHARACTERISTICS OF THE INTRUSIONS

Burgeo Intrusive Suite

The Burgeo intrusive suite displays a broad and continuous compositional variation, with a range in silica values from 45.5 to 77.8 weight percent (Table 1; Figure 2). This variation generally follows an evolutionary trend with early gabbroic rocks intruded by porphyritic biotite granites and late biotite–muscovite and high-silica granites. Major- and trace-element variations within the entire suite follow the expected differentiation trends: there is a steady decrease in FeO, MgO, CaO, Sr, Zr, F and an increase in K_2O, Rb and U (see Figure 3 for examples; also Table 2). The agpaitic index [mol.$(Na_2O+K_2O)\div Al_2O_3$] is always less than 1 indicating that the suite is peraluminous to metaluminous. The variation in the AFM diagram (Figure 4) shows a tight coherence possibly indicating petrogenetic relationships between the various units.

Chetwynd Granite

The Chetwynd Granite is characterized by a high mean silica value and ranges from 62.5 to 78.1 wt. percent (Table 1). This range is reflected in the wide variation of values for

Table 1. Statistics for major oxides (in weight %) and trace elements (in grams per tonne) in the Burgeo intrusive suite, Chetwynd and François Granites.

	Burgeo			Chetwynd			François		
	Mean	Max.	Min.	Mean	Max.	Min.	Mean	Max.	Min.
SiO_2	67.7	77.8	45.5	73.6	62.5	78.1	74.5	79.8	69.1
Al_2O_3	15.1	20.5	11.7	13.7	17.0	12.0	13.5	15.4	10.5
Fe_2O_3	1.19	5.19	.01	.75	1.75	.12	.51	1.61	.04
FeO	2.19	7.97	.01	.62	3.91	.01	.68	2.31	.01
MgO	1.30	13.90	.03	.38	3.08	.03	.21	.76	.02
CaO	2.30	9.56	.07	.81	3.23	.08	.61	1.83	.09
Na_2O	3.35	5.72	1.50	3.55	4.48	2.89	3.71	6.34	2.90
K_2O	4.62	9.29	.33	5.72	8.11	3.88	5.23	7.03	3.32
TiO_2	.59	1.96	.03	.30	.82	.05	.17	.51	.03
MnO	.08	.22	.01	.04	.14	.01	.05	.11	.01
P_2O_5	.19	.66	.01	.06	.27	.01	.04	.14	.01
L.O.I.	.89	3.89	.32	.55	.37	.13	.60	1.49	.32
	Mean	Max.	Min.	Mean	Max.	Min.	Mean	Max.	Min.
Li	42	166	1	21	51	6	52	129	9
Be	4	22	1	3	6	1	6	69	2
F	595	2616	14	402	1142	35	492	1937	24
V	64	470	1	17	148	2	5	7	1
Cr	28	309	1	14	164	1	9	1	38
Ni	9	377	1	3	46	1	1	8	1
Cu	13	562	1	4	23	2	3	45	1
Zn	52	130	5	32	71	7	33	78	5
Ga	21	101	8	29	91	17	21	33	11
Rb	190	512	4	229	367	131	388	1131	144
Sr	200	531	4	74	203	1	72	287	3
Y	33	122	1	49	133	7	51	241	6
Zr	174*	642	10	157	423	25	106	312	21
Nb	15	44	1	20	43	4	29	97	8
Mo	3	61	1	3	21	1	2	8	1
Ag	.1	.3	.1	.1	.3	.1	.1	.2	.1
Sn	1	40	1	1	8	1	1	13	1
Ba	648	3775	5	278	823	6	200	729	1
La	34	149	1	45	124	5	33	158	1
Ce	82	205	1	112	271	8	86	294	1
W	1	18	1	2	8	1	1.4	22	1
Pb	23	61	2	26	39	7	32	81	4
Th	18	99	0	28	63	1	42	89	8
U	3.6	41.5	1	3.7	9.4	.1	10.3	41.7	1.7

* n = 255 (samples from NTS sheets 11P/10 and 11 only)

CaO, MgO, FeO and Al_2O_3. Trace elements such as Sr, V, Zr and Y also display a wide range of values (Table 1). The lower silica granites are the K-feldspar-megacrystic variety containing cumulate feldspar, which clearly increases the proportions of K_2O, Na_2O, CaO and Al_2O_3, and decreases the proportion of SiO_2, in the rock (Table 2).

The higher silica granites are strongly depleted in F, Sr, MgO and CaO, and K_2O also shows a negative correlation with SiO_2. Rb/Sr ratios increase strongly with a small increase in SiO_2, although Rb only doubles in abundance (Figure 3; Table 2). These variations are reflected by a decrease in modal biotite and feldspar. Incompatible elements such as U and

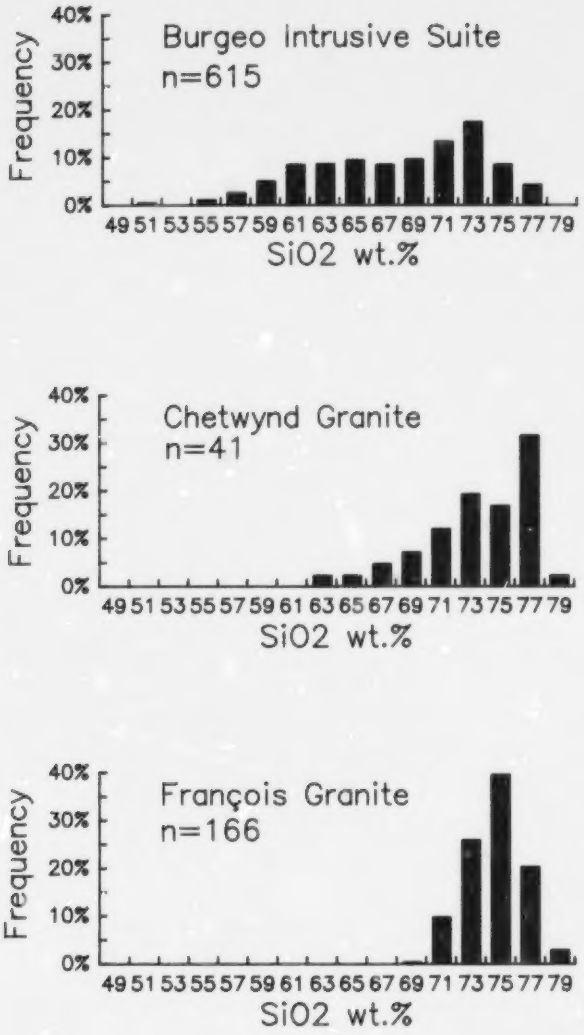

Figure 2. *Plot of SiO_2 frequency distributions for the Burgeo intrusive suite, Chetwynd and François granites.*

Th, and granophile economic elements such as Mo, Sn, W and Cu, show no significant enrichment with SiO_2.

François Granite

The François granite is lithologically similar to the Chetwynd Granite, but is characterized by more extreme depletion and enrichment of trace elements. It shows a more restricted silica range, higher mean silica, greater depletion of certain components such as CaO, MgO, Cu and Sr, and stronger enrichment of some incompatible elements, e.g., U, Th and Rb (Table 1; Figure 3). Variation diagrams show a pronounced increase in Rb/Sr ratios, F and U, and only small changes in SiO_2. Enrichment and depletion are greatest in the late, leucocratic, aplite (Unit Df_5), which occurs as dykes, and forms the core of the northeastern lobe of the granite (Unit Df_9) (Table 2). The relatively mafic granodiorite (Units Df_4 and Df_8) is also lowest in silica and consequently contains the highest CaO, MgO, FeO and Sr values and the lowest Rb, U and Th values (Table 2). These variations are reflected in airborne radiometric maps of Broome *et al.* (1987), which clearly show that the François granite is distinctly radiogenic compared with the adjacent granites. This is also seen in Table 1 and Figure 3, which show that the François granite contains the highest U, Th and Pb values. Isolated anomalous Cu, Sn and W values occur, but mean values are not significantly enriched.

DISCUSSION

Petrogenesis

Most granitoid units discussed above contain an abundance of K_2O and inherited zircons indicative of a continental-crust component in the magmas. The initial $^{87}Sr/^{86}Sr$ ratio of 0.7092 obtained by Higgins (1980) is also indicative of a significant continental-crust component in the granites. However, the gabbroic units reflect a mantle-derived component. The variety of rock types within the Burgeo intrusive suite is similar to calc-alkaline suites such as the Sierra Nevada Batholith (Bateman and Chappell, 1979) or the Coastal Batholith of Peru (Cobbing and Pitcher, 1972) and adjacent complexes (e.g., Kontak *et al.*, 1984; see Figure 4). However, in the Coastal Batholith of Peru, tonalite is the dominant lithology having subsequent intrusion of lesser amounts of granite. This evolutionary trend is typical of calc-alkaline complexes developed at the outer edge of continental magmatic arcs. The abundance of relatively potassic granite found in the Burgeo intrusive suite is unlike that found in calc-alkaline suites such as the Sierra Nevada. Another granite-dominated suite, such as that of southwest England, which contains abundant K-feldspar-porphyritic granite, may provide a closer comparison for the K-feldspar-rich units of the Burgeo intrusive suite.

Studies on the Coastal Batholith of Peru (see Harmon and Barreiro, 1984) support a subduction-related origin for most of the Coastal Batholith associated with extensive fractional crystallization of early tonalite to produce the later granodiorite and granite plutons. Contamination and/or partial melting of continental crust is considered to be a minor component in the Peruvian granitoids, even though the plutons were intruded into geosynclinal metasediments, e.g., Atherton (1984). A similar evolution is proposed for the granites from southern Chile by Bartholemew and Tarney (1984). They conclude, however, that the generation of large masses of potassic granite, 25-km wide and perhaps several hundred kilometres in lateral extent, remains a problem.

A lead isotope study of plutonic rocks from the Western and Coastal Cordilleras, in Peru, by Mukasa and Tilton (1984), showed a spatial association of basement lead in the Phanerozoic plutonic rocks and exposed Precambrian granulitic gneisses. The more felsic rocks appeared to contain an old crustal Pb component interpreted to be the result of high-level mixing of gneisses with mantle-derived material.

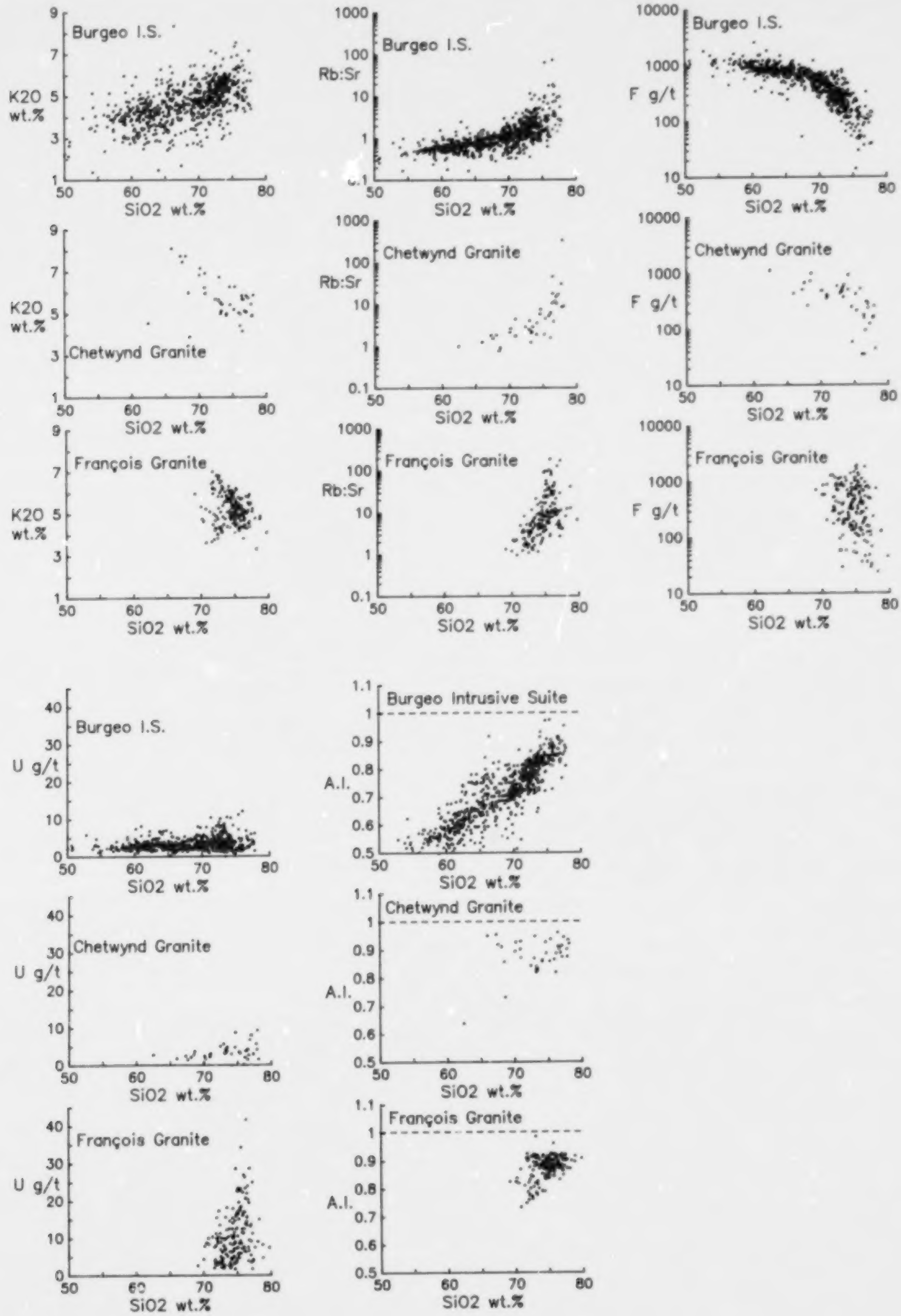

Figure 3. *Harker variation diagrams for the Burgeo intrusive suite, Chetwynd and François granites.*

Table 2. Analyses of representative samples from the Burgeo intrusive suite, Chetwynd and François granites. Major elements in weight percent; trace elements in grams per tonne.

	Burgeo Intrusive Suite				Chetwynd Granite		François Granite			
Unit	Sb_1	Sb_1	Sb_1	Sb_2	Dc	Dc	Df_2	Df_6	Df_8	Df_9
Rock	tnlt	grnt	grnt	grnt	grnt	grnt	grnt	grnt	grdr	grnt
type+	a	b	c	d	e	f	g	h	i	j
SiO_2	65.6	59.7	76.4	71.55	70.95	77.85	76.65	74.45	71.80	75.90
Al_2O_3	15.42	16.65	13.09	13.79	14.91	12.03	12.10	13.76	14.82	13.27
Fe_2O_3	1.37	1.30	.54	.70	.79	.58	.32	.33	.44	.30
FeO	2.74	4.78	.23	1.04	.84	.21	.78	.85	1.15	.54
MgO	1.47	2.96	.15	.47	.52	.15	.12	.16	.50	.08
CaO	2.87	4.38	.35	1.22	1.04	.52	.37	.78	1.25	.16
Na_2O	3.36	3.03	3.02	3.43	3.97	3.19	3.35	3.27	4.13	4.29
K_2O	4.35	4.40	6.41	5.20	5.96	4.87	5.08	6.12	4.49	4.37
TiO_2	.73	1.08	.13	.32	.35	.16	.15	.17	.26	.09
MnO	.10	.13	.03	.06	.05	.03	.04	.04	.07	.04
P_2O_5	.24	.39	.03	.10	.07	.03	.02	.04	.09	.02
L.O.I.	1.08	1.03	.52	.72	.82	.57	.49	.39	.77	.75
Li	117	36	28	46	26	18	29	43	92	44
Be	.	.	3	.	2	3	4	4	7	6
F	696	.	92	.	390	169	159	348	530	282
V	91	127	6	39	30	7	1	1	14	1
Cr	26	72	9	5	21	12	7	6	8	5
Ni	9	25	1	1	2	1	1	1	2	1
Cu	8	23	1	2	3	3	1	1	4	10
Zn	50	81	18	44	32	12	31	29	52	30
Ga	19	18	17	15	22	20	19	20	26	29
Rb	185	184	313	89	236	264	206	318	362	762
Sr	189	320	39	106	137	31	19	75	217	23
Y	46	38	12	20	60	20	39	37	62	141
Zr	368	.	67	279	144	57	102	113	123	69
Nb	17	17	9	17	16	16	21	15	32	59
Mo	3	4	2	3	2	2	2	2	2	2
Ag	.1	.	.1	.1	.1	.1	.1	.1	.1	.1
Sn	.	.	1	.	2	1	1	1	1	2
Ba	648	955	82	357	510	54	72	243	585	52
La	35	23	12	53	24	17	36	35	6	11
Ce	74	57	9	102	91	51	110	124	38	39
W	.	.	1	.	2	1	1	1	1	2
Pb	18	14	28	28	25	27	18	23	30	48
Th	17	.	45	24	19	28	31	40	35	48
U	3.1	.	3.6	4.5	3.1	4.2	3.9	4.9	8.0	16.9

+ a: tonalite #1941337; b: mafic porphyritic granite #1940444; c: high-silica granite #2242002; d: biotite−muscovite granite #1941202; e: coarse granite #1940934; f: medium granite #1940693; g: coarse equigranular granite #2242034; h: equigranular granite #2242035; i: porphyritic biotite granodiorite #2242023; j: medium equigranular granite #2242018

The hornblende-bearing granodioritic, tonalitic and gabbroic intrusions of the Burgeo intrusive suite may likewise represent a mantle-derived component. The biotite ± muscovite granodiorites and granites in the Burgeo intrusive suite, and their common association with migmatitic metasediments, may be the result of high-level melting of continental crust induced by emplacement of more mafic magmas. Dunning (*in* O'Brien *et al.*, 1986) notes that the results of isotopic studies, on the adjacent Silurian La Poile Group, indicate that either Grenville-aged continental crust or sediment derived from it was involved in melting to produce the La Poile magma. A similar source would also

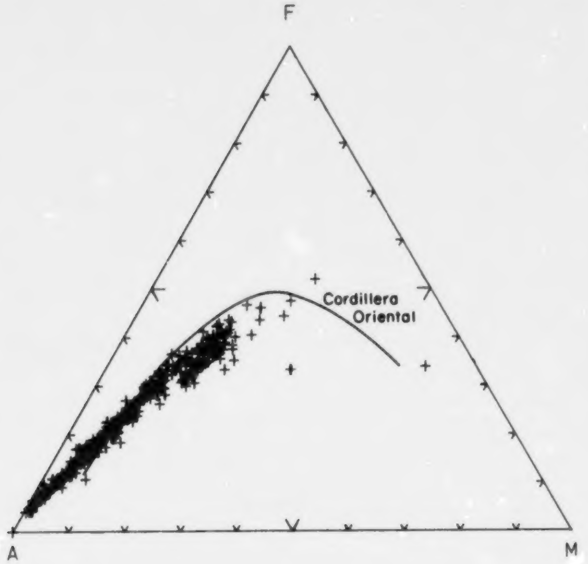

Figure 4. *Plot of total alkalis ($Na_2O+K_2O = A$):(total iron as $FeO = F$):($MgO=M$) for the Burgeo intrusive suite; trend shown is for calc-alkaline plutons of the Cordillera Oriental, Peru (from Kontak et al., 1984).*

be available for the later Burgeo intrusive suite granites. Using the classification of Pearce *et al.* (1984), the Burgeo intrusive suite should display geochemical features of volcanic-arc granites and/or syn-collision granites. A plot of Rb versus Y+Nb (Figure 5) shows that the majority of samples from the Burgeo intrusive suite fall in the volcanic-arc and within-plate granite fields. However, many samples lie in the within-plate field, usually represented by alkaline and peralkaline granites, indicating that the field boundaries are not precise.

The Chetwynd and François granites are high-silica, highly differentiated, two-feldspar, biotite granites that lack basic and intermediate phases (Figures 2 and 3). Both granites were intruded to high structural levels and, in the case of the François granite, ring complexes were developed. Late-stage crystallization of water-saturated melts is indicated by the presence of miarolitic cavities, tuffisites and granophyre. The primary magma was undersaturated in water as shown by its ability to rise through the crust and probably had only about 10 percent crystals as indicated by rhyolitic porphyry dykes in the country rocks. Similarly, the eastern margin of the northeastern complex of the François granite is a 20-m-wide, chilled zone with coarse biotite, quartz, alkali feldspar and plagioclase phenocrysts set in an aplitic matrix.

The François granite is particularly enriched in Rb, U and Th compared with the Burgeo intrusive suite (Figure 3; Table 1). Both granites are strongly depleted in FeO, MgO, CaO, Cu and Sr (Table 1). In the François granite, the late emplacement of plagioclase–porphyritic granite indicates that fractional crystallization of plagioclase occurred in both ring complexes. This is also shown by the strong depletion of Sr

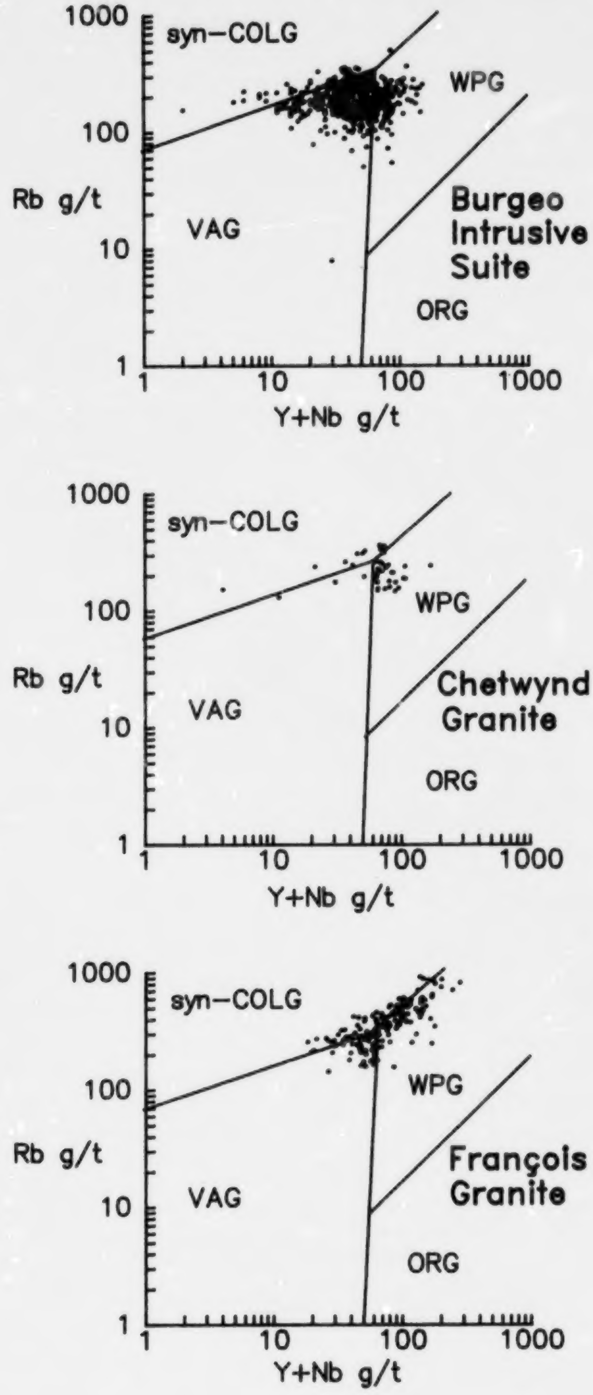

Figure 5. *Plot of Rb versus Y+Nb for the Burgeo intrusive suite, François and Chetwynd granites; fields for "volcanic arc granites" (VAG), "ocean ridge granites" (ORG), "syn-collisional granites" (syn-colg), and "within plate granites" (WPG) from Pearce et al. (1984).*

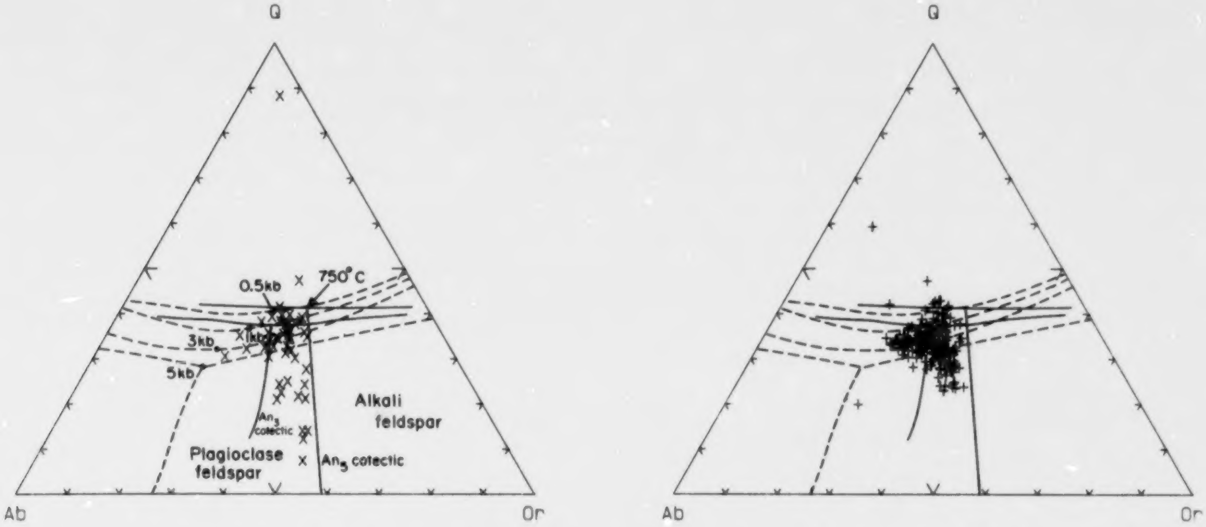

Figure 6. *Plot of normative Q:Ab:Or ratios for the Chetwynd and François granites with cotectics from Tuttle and Bowen (1958) and An_5 from James and Hamilton (1969).*

and CaO in the subsequent intrusion, which forms the core of the northeastern ring complex and has Rb/Sr ratios in excess of 100 (see Figure 3). The François granite contains rare, biotite-rich, cumulate layers indicating that fractionation of biotite could be the cause of the strong depletion of FeO and MgO. The depletion trend shown by K_2O indicates that K-feldspar was also fractionated out during crystallization.

Some units of the Burgeo intrusive suite, in the Bay de Vieux area, are composed of undeformed, biotite-poor granite similar to the Chetwynd and François granites and may be about the same age. Thus the Chetwynd and François granites may represent the final episode of Burgeo plutonism.

The low CaO content and therefore normative anorthite, of the Chetwynd and François granites, allows the use of the normative quartz–albite–orthoclase diagrams of Tuttle and Bowen (1958) and James and Hamilton (1969) to give an estimate of P_{H_2O}–T conditions for crystallization of the magmas. Figure 6 shows that the François and Chetwynd granite compositions cluster around the ternary minimum of 725°C, in the water-saturated granite system at 1 kilobar P_{H_2O}. An origin by fractional crystallization of a non-minimum melt, parent magma is indicated for the Chetwynd Granite by the trend along the Ab–Or cotectic in a system with 3 to 5 percent An at 1 kilobar P_{H_2O} (James and Hamilton, 1969) and suggests that crystallization occurred in the temperature range 875 to 750°C. The lack of a trend toward the Ab–Or join in the François granite indicates that the magmas crystallized at a lower temperature and the magma composition was close to a minimum melt.

The Chetwynd and François granites are similar to those of southwest England being high-level, alkali feldspar-porphyritic biotite granites. The only significant differences are the abundance of Sn–W mineralization and apparent abundance of primary muscovite in the southwest England granites. These may reflect differences in source rocks. Pearce *et al.* (1984) have indicated that the southwest England granites, although undeformed, belong to the syn-collisional granite group. Using the Rb versus Y+Nb plot (Figure 5) it is clear that the François and Chetwynd granites, do not lie within the syn-collisional granite field. However, the chronology and geological setting of these granites indicate that they are the product of a major orogeny.

Relationship of Intrusion to the Development of Major Structures

Minimum ages of movement on several of the major faults in the Hermitage Flexure area are given by cross-cutting relationships of dated granites and, in one case, mineralized veins. The Chetwynd Granite is described by Chorlton (1980b) as brecciated by the latest movements of the Bay d'Est Fault, indicating that the latest movements are post-377 Ma. The Dragon Bay Fault zone is cut by the François granite. Although undated, the François granite is similar to the Middle Devonian (approximately 367 Ma; Kontak *et al.* 1988) Ackley Granite and to the Chetwynd Granite. These correlations would imply that movement on the fault ceased by the Middle Devonian. Similarly, the fault contact with the Grey River enclave and associated shear zones within the Burgeo intrusive suite are truncated by the François granite. At Grey River, wolframite-bearing quartz veins cut mylonitized Burgeo granitoids and the Grey River enclave metamorphic rocks. Higgins (1980) reports a series of K/Ar (muscovite) ages from the veins, which indicate a long period of vein emplacement from 386 Ma to 329 Ma. These concentrate in the range 360–370 Ma, which could possibly include the age of the François granite. A sample of shear zone granite gave a K/Ar (biotite) age of 395 ± 8 Ma.

Mineralization and Exploration Potential

Known metallic mineralization associated with the Burgeo intrusive suite, and the Chetwynd and François granites is of limited extent. Wolframite-bearing quartz vein mineralization was discovered in 1955 (Bahyrycz, 1956) in the metamorphic rocks at Grey River and in the adjacent granite of the Burgeo intrusive suite. Associated base metals include molybdenite, chalcopyrite, galena and sphalerite. Surface drilling and extensive underground exploration were carried out and high-grade mineralization was discovered. Numerous reports are available on the geology and metallurgy of the tungsten deposits, e.g. Bahyrycz (1956, 1957); Higgins (1980, 1985); and Mathieu (1961). These indicate that high-grade tungsten ore is present in an extensive quartz vein system with the major veins extending for 1 to 2 km.

Following the discovery of the Grey River tungsten prospect, prospecting by the Buchans Mining Company in the upper Grey River area, during the late 1950's, resulted in the discovery of numerous, tungsten and pyrite showings in the vicinity of the contact of the Burgeo intrusive suite and the Bay du Nord Group.

Along Dog Cove Brook, about 5 km north of the main Grey River tungsten vein system, Bahyrycz (1956) reported widespread mineralized quartz veins in a fault zone. The veins carried Cu, Pb, Zn and Fe sulphides and assayed anomalous gold and silver. The highest values were obtained from just west of 'Beaver Pond', where galena–sphalerite–pyrite mineralization occurs in a quartz vein enclosed by altered granite. This was subsequently confirmed by Thurlow (1981). The altered granite wallrock assayed 2.90 oz/ton Au over 1 ft (Bahyrycz, 1956) and 0.74 oz/ton Au over 1.83 ft (Thurlow, 1981).

Molybdenite mineralization occurs in the Burgeo intrusive suite about 16 km north of Burgeo (O'Brien, 1983). The mineralization occurs as disseminations in widespread aplite veins.

McConnell (1985) reported anomalous Sn, Pb and W values from the François granite but mineralization has not yet been discovered. The François granite is anomalously radioactive and contains distinctly higher F, U, Th and Pb values than the Burgeo intrusive suite and the Chetwynd Granite. The François granite is extensively altered in places with kaolinization of feldspars and black quartz crystals. This is perhaps a result of radiation damage. The locally high fluorine content of biotite-poor granites indicates that fluorine was concentrated in the system and hydrothermal alteration is locally present. McConnell (1985) noted that topaz was discovered in a loose block in the Devil Bay area, and this is clearly indicative of hydrothermal mineralization.

Highly differentiated, fluorine-rich parts of the Chetwynd Granite have been identified (O'Brien, unpublished data) and these have potential for granophile mineralization. The spatial relationship, on a regional scale, between gold mineralization at Hope Brook (e.g. O'Brien, 1987, 1988) and the Chetwynd Granite may reflect some genetic relation between the two, and perhaps highlight the importance of the late high-silica granites in the metallogenic evolution of the region.

The metasedimentary and associated rocks, east of Grey River, contain an extensive silica deposit, which has been interpreted as either a vein or a metasediment. This deposit has been assessed by Butler and Greene (1976) who report a proven tonnage of 12 million short tons (approximately 11 million tonnes) averaging 95.5 percent SiO_2 and 1.9 percent Al_2O_3 with selected portions containing higher silica and lower alumina values.

There is only minor hydrothermal alteration associated with the major shear zones. Sericitization and chloritization are common along the faults. The Bay de Loup Brook Fault, 10 km north of Bay de Loup, is extensively silicified with milky quartz and trace amounts of pyrite cementing fault-brecciated granite.

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GOLD MINERALIZATION AND IGNEOUS ACTIVITY IN RELATION TO SYNMETAMORPHIC FOLDING, THRUSTING AND WRENCHING: A GENERAL MODEL FOR THE SOUTHEASTERN MARGIN OF THE CANADIAN APPALACHIANS

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Newfoundland Mapping Section

Introduction

An important class of Atlantic Canadian gold deposits, including BP-Selco's Hope Brook Mine in Newfoundland, Western Mining's Forest Hill and Beaver Dam mines in Nova Scotia and the Gordex Mine at Cape Spencer, New Brunswick, occur near regional lineaments in Paleozoic rocks of the Appalachian Orogen. Most of these lineaments reflect ductile fault systems or tightly braided mylonite zones, the most significant of which record a protracted history of displacement and yield evidence of persistently high-heat flow. Economically significant lineaments are characterized by (1) their common location above Precambrian or Cambro–Ordovician ancestral structures, (2) repeated volcanism and plutonism along the fault lineament, (3) polymetamorphism of some or all of the gold-bearing rocks near the fault lineament, and (4) post-Devonian fault reactivation and lineament enhancement.

Regional Setting of Lineaments

Within the accreted, southeastern margin of the Appalachian Orogen, tectonostratigraphic terranes are commonly bound by large-scale transcurrent faults or dextral megashears that delimit wrench-fault systems varying between 10^3 to 10^5 km^2 in area. The principal shear planes or east-trending master faults initially formed at a relatively late stage in Acadian (Siluro–Devonian) regional deformation; however, many developed near older faults or were ductilely reactivated by Hercynian (Carboniferous) movements. Orogenic flexures or megakinks were produced where wrench-generated, northeast-trending, Acadian-aged folds, foliations and thrust faults were rotated, along with pre-existing structures, toward the ductile master faults. In Appalachian terranes, governed by either Acadian or Hercynian wrench-faulting, the regionally predominant periclinal folds, and slaty to spaced foliations are upright. However, narrow transitions to recumbent structures occur where the regional strain increases as the suprastructure of the orogen passes gradationally into its infrastructure or abruptly into its basement. Inhomogeneous shortening of the Acadian suprastructure was accommodated by fold-related thrusts, which formed during the development of low-grade regional foliation. Although platy-foliated, syntectonically-veined or mylonitic rocks are locally gold-bearing, the amount of stratigraphical separation across ductile faults of Acadian or Hercynian age is commonly inconsequential. Therefore, despite the fact that these faults affect suprastructure, infrastructure and basement, they are probably not deep structures, although they may be sited along ancestral fault zones that root in the lower crust or mantle.

Types of Gold Mineralization

Gold deposits in the Appalachian Meguma, Avalon, and Gander terranes occur preferentially in the suprastructure of the fold belt and are tectonically localized in slate belts composed of variable amounts of sedimentary and volcanic rocks. Less commonly, gneiss and schist belts near the infrastructure–suprastructure transition, or near basement-cover contacts, contain gold mineralization. The slate belts host two end-member types of mineralization: (1) nugget gold in Bendigo-type, saddle-reef, quartz–carbonate lodes (e.g., those in sandy and shaly flysch of the exotic Meguma Terrane of mainland Nova Scotia), and (2) disseminated gold in Carolina-type, high-alumina, high-silica, sericitic schist zones (e.g., those in pyroclastic and epiclastic deposits of the Gander (?) Terrane of southwest Newfoundland). The stratigraphy, terrane-type, age and composition of the mineralized country rocks are largely incidental.

Origin and Nature of Fluids

In many, although not in all gold districts, mineralization and alteration occur in association with igneous activity. Mineralizing fluids presumably originated from, or were heated by, granodiorite–tonalite plutons and satellite porphyries. In places, these granitoids are spatially associated with H_2O-, CO_2- and P-rich gabbro–diorite complexes and mafic dyke swarms, the loci of which are regularly spaced along some fault lineaments. All of these intrusive rocks probably formed in, or passed through, the deep basement and/or infrastructure of the orogen and ascended in sequence at various stages of regional deformation. Those of Acadian age were emplaced into the slate belts during high-T, low-P, dynamothermal metamorphism. Pseudoporphyroblasts of sulphides, carbonates, aluminosilicates and phosphates (some containing rotated inclusion trails), record the unique physicochemical conditions locally prevalent during periods of fluid-controlled, synmetamorphic, synplutonic alteration. Regional metamorphism at the time of emplacement of the gold-generating, felsic–mafic, composite intrusions and gold-

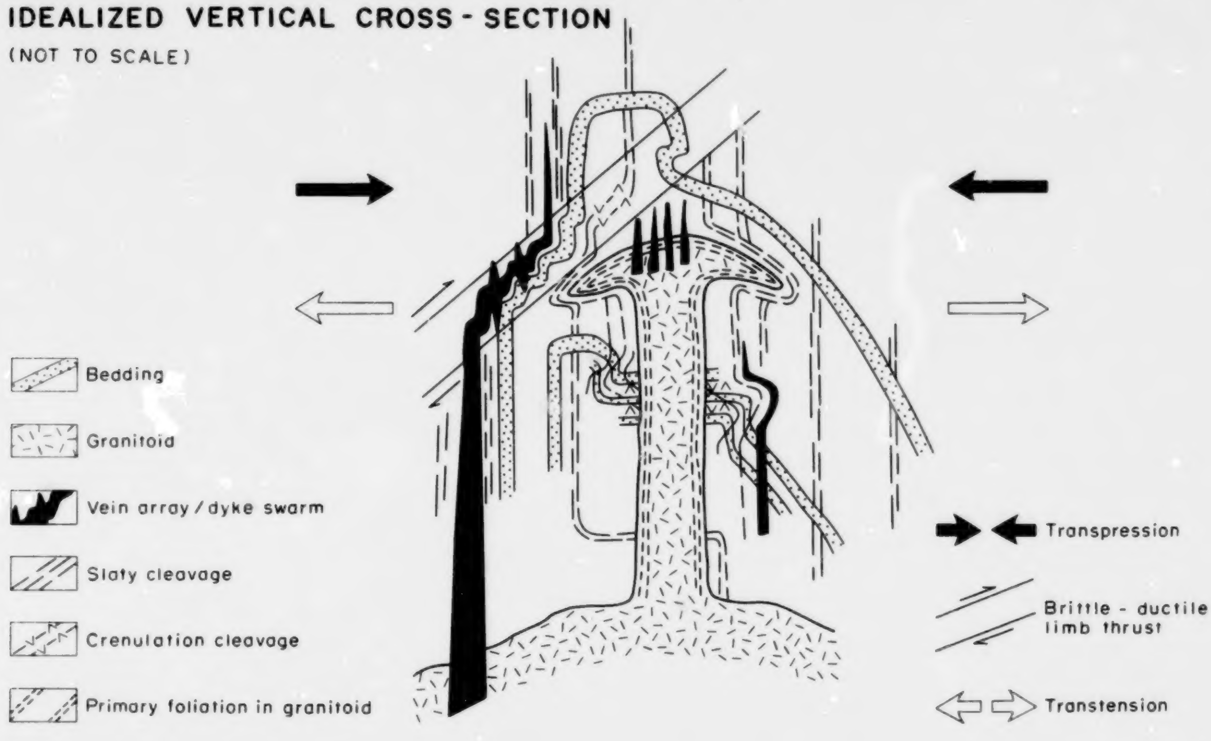

Figure 1. *Illustration of some geometric relationships between fold and thrust related foliations, and auriferous veins and intrusions. Note the direction of transpressional and transtensional forces.*

bearing, quartz–carbonate, crack-seal, vein arrays was generally in the greenschist facies, although some host rocks were metamorphosed in the amphibolite facies before or after the introduction of magma and fluids.

Deformation Kinematics and Dynamics near Gold Deposits

Viscoelastic deformation near gold deposits was accomplished by two complementary processes: (1) dilation and synchronous shortening of intrusive bodies and veins in localized parts of fault zones or fold-hinge zones, and (2) a more widespread extension and flattening of the country rocks through the progressive development of coaxial folds and fabrics.

Zones of large total strain are host to a greater variety and a larger number of syndynamic, synkinematic intrusions than zones that were never appreciably strained. Regardless of whether these zones are vertical or horizontal, quartz–carbonate veins and mafic -felsic sills and dykes are emplaced along and, to a lesser extent, across the foliation displaying the maximum anisotropy. Variably altered, mineralized, aged and foliated sheet-like plutons and sills occupy foliation-parallel shear fractures in the ductile fault zones at the expense of their conjugate, cross-strike pair. Whereas these rocks invariably suffer some of the strong regional deformation affecting the slate belts, gold veins and hypabyssal intrusions in the negligibly strained, simply folded areas between faults, illustrate two distinct deformation styles. An autokinematic style of deformation affects dykes displaying syn-intrusion offset of host-rock markers. Deformation of crystallizing magma between the walls of conjugate pairs of dilating dykes produced sigmoidal internal foliations that, in the case of one set, are perpendicular to the regional cleavage. In contrast, gold veins containing cleavage inclusions and laminated, fibrous or vuggy crystals are deformed by several orders and orientations of folds, and occupy systematic fractures parallel to the AB-, AC- and BC-kinematic planes of regional periclines.

Roles of Transtension and Transpression in Localizing Gold Deposits

Regional systems of interrelated folds, transcurrent faults and thrust faults (reactivated by normal fault movements) occur along the southeastern margin of the Canadian Appalachians. Within most of the known goldfields on this margin of the orogen, particular structures are identifiable that act as conduits and provide traps for gold-bearing fluids, in much the same way that similar tectonic features localize oilfields in southern California. Contemporaneous and cyclic, brittle–ductile episodes of horizontal extension (transtension) and horizontal shortening (transpression) were operative throughout these systems, but their effect on intrusions of all scales has special ramifications for the localization of gold deposits. Orthogonal and oblique extensions across developing foliations controlled the geometrical shapes, the growth textures and some of the deformation features of the gold quartz veins and variably altered hypabyssal intrusions. For example, in upright folds or faults invaded by vertical syntectonic intrusions, country-rock layering as well as foliation-parallel veins and dykes were bent into flat orientations by extension and then rebuckled into steep attitudes by compression as progressive regional deformation focused on particular structures (see Figure 1). Since these processes were continuous, some individuals in the vein arrays and dyke swarms were generated while others were being folded or boudinaged. In places, granitoid plutons rose along steeply dipping faults or upright fold axial surfaces and spread laterally where these structures became flat-lying or recumbent, or where the roofs of plutons behaved diapirically. Here, ductile horizontal stretching locally produced flat fabrics in the granitoids and distorted the once upright cleavage fans. It also caused the roofs of the granitoids and the adjacent parts of the fold-hinge zones or thrust sheets to fail in extension, as space was created for vertical satellite intrusions emanating from the plutons. Pulses of mineralization, alteration and igneous activity occurred during periods of extension intimately associated with fold and foliation development and fault movement.

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SUMMARY OF THE GEOLOGY BETWEEN LA POILE BAY AND COUTEAU BAY (11O/9 AND 11O/16), SOUTHWESTERN NEWFOUNDLAND

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ABSTRACT

The geology of the La Poile Bay–Couteau Bay region is summarized by describing the evolution of the Bay d'Est, Cinq Cerf and Grand Bruit fault zones of the Hermitage Flexure system. Southeast-dipping listric thrusts or fold slides, important constituents of these fault zones, juxtapose Precambrian, Precambrian-Cambrian, Ordovician and Silurian rocks, which are separately disposed in four thrust sheets. Phyllite, schist and gneiss units of variable age, but of predominantly sedimentary and volcanic origin, were translated northwestward and dynamothermally metamorphosed in the greenschist facies during Silurian D_1 deformation. The thrusting of Precambrian schist–gneiss units over Precambrian–Cambrian phyllite units is thought, however, to have occurred during pre-Silurian movements in the Grand Bruit fault zone. An unconformity at the base of the Silurian La Poile Group is interpreted to be displaced by thrusting along an old fault structure in the Cinq Cerf fault zone.

The Siluro–Devonian D_2 deformation affected parts of each of the four thrust sheets after their assembly but prior to the cessation of regional metamorphism. D_2 deformation is best developed within the Silurian plutons and batholiths, and in country rocks marginal to these intrusions. Syn-D_2 extension of the thrust sheets facilitated the sheet-like emplacement of Silurian magmas of alternating felsic and mafic composition. It was also responsible for reorientating and then refolding D_1 or older structures and permitted, at least within the thrust sheet containing the La Poile Group, the infiltration of auriferous fluids along pre-existing foliations, folds and faults.

INTRODUCTION

The study area is located in the Central Mobile Belt of the Newfoundland Appalachians and forms part of the western Hermitage Flexure (Figure 1). It consists of several distinct structural and lithological assemblages, some of which may correlate with the Avalon and Dunnage zones. The region between La Poile Bay and Couteau Bay in southwestern Newfoundland comprises the Grand Bruit–Cinq Cerf map area surveyed in 1986 (O'Brien, 1987) and the La Poile Bay–Roti Bay map area surveyed in 1987 (O'Brien, 1988).

The purpose of this report is to present 1:100,000 scale compilation maps and 1:50,000 scale sections of the study area showing the regional disposition of lithological units and major structures. In addition, the geological history of the region is summarized from the perspective of local faults within the Hermitage Flexure system, and new correlations and age assignments of lithostratigraphic units and tectonic structures are suggested. In this paper, the names of informal map units are distinguished by the use of upper case to identify place and lower case to distinguish representative lithology.

GENERAL STATEMENT

The region between La Poile Bay and Couteau Bay is underlain by stratified and intrusive rocks ranging from the Precambrian to the Devonian (Figure 2; Cooper, 1954; Chorlton and Dallmeyer, 1986; Dunning *et al.*, 1988; Dunning and O'Brien, *in press)*. The regional distribution

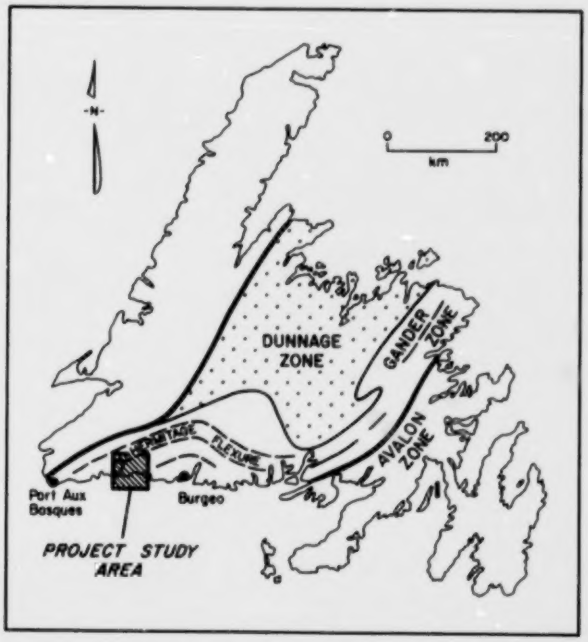

Figure 1. *Location map showing the study area in relation to the Hermitage Flexure and the Avalon, Gander and Dunnage Zones.*

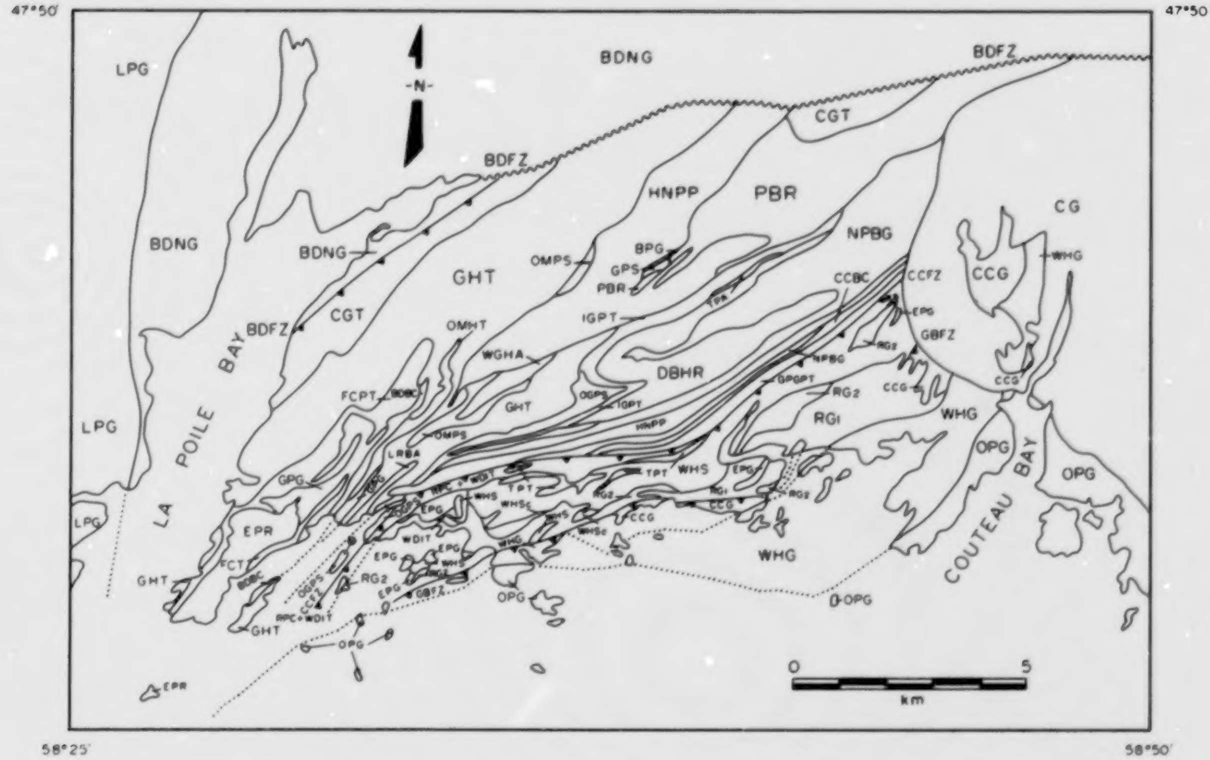
47°50'
47°50'
-N-
LPG
BDNG
BDFZ
CGT
HNPP
PBR
CG
NPBG
BDNG
GHT
OMPS
BPG
GPS
PBR
CCBC
CCFZ
CCG
WHG
CGT
BDFZ
OMHT
IGPT
EPG
GBFZ
RG2
WGHA
DBHR
NPBG
GPGPT
RG2
CCG
FCPT
GHT
OGPS
IGPT
HNPP
RG1
WHG
CCG
LPG
LA POILE BAY
COUTEAU BAY
OMPS
LRBA
TPT
WHS
EPG
OPG
OPG
GPG
LPG
EPR
WDIT
WHS
WHSc
CCG
FCT
WHG
EPG
WHG
GHT
RG2
OGPS
CCFZ
EPG
GBFZ
OPG
OPG
RPC+WDIT
GHT
OPG
EPR
0
5
km
58°25'
58°50'

Figure 2. *Geological map of the region between La Poile Bay and Couteau Bay, southwestern Newfoundland (1:100,000 scale). Location of the Hope Brook Gold Mine is indicated.*

LEGEND

DEVONIAN

Plutonic Rocks

CG	*Chetwynd Granite: pink, equigranular biotite, granite; microspherulitic dykes*

SILURIAN
La Poile Group

Cross Gulch tuff

CGT	*lithic tuff composed of a variety of volcanic fragments, minor argillite and sandstone*

Gallyboy Harbour tuff

GHT	*pink, medium and coarse, locally layered and stratified, quartz–feldspar crystal tuff; rare blocks of volcanic and sedimentary rocks*

Flint Cliff Pond tuff

FCPT	*buff-weathered lithic tuff; coarse agglomerate composed of massive, pink rhyolite blocks; minor siliciclastic beds and rare lava flows*

Eastern Point rhyolite

EPR	*massive, pink, aphanitic lava; locally welded*

(Legend continued on next page...)

LEGEND

(Continued from previous page)

French Cove tuff	FCT	*well-stratified, lithic tuff and breccia containing sedimentary and volcanic fragments; minor lithic–crystal tuff*
Guiders Pond grit	GPG	*poorly sorted, grey-brown, locally stratified, quartz–feldspar grit and tuffaceous wacke; minor boulder conglomerate and olistostrome; rare slate*
Black Duck Brook conglomerate	BDBC	*polymictic, cobble and boulder conglomerate; siliciclastic sandstone interbedded with laminated argillite; minor grey slate*
Old Man Hill tuff	OMHT	*buff-weathered, fine grained, lithic tuff composed of pink rhyolite fragments*
Withy Gulch Hill agglomerate	WGHA	*agglomerate and breccia composed of massive, flow-banded and flow-folded rhyolite; minor pink and cream rhyolite*
Old Man Pond slate	OMPS	*grey slate and laminated argillite; minor thin-bedded sandstone and lithic tuff*
Little Roti Bay agglomerate	LRBA	*crudely stratified, coarse agglomerate and breccia containing polymictic volcanic blocks; purple lithic tuff*
Outside Gull Pond slate	OGPS	*grey slate and laminated argillite; rare thin-bedded sandstone*
Georges Pond slate	GPS	*grey slate and laminated argillite*
Butterfly Pond grit	BPG	*poorly sorted, grey-brown, quartz–feldspar grit and tuffaceous wacke; grey slate*
Phillips Brook rhyolite	PBR	*massive and finely laminated, cream and pink lava; breccia composed of flow-banded rhyolite*
Twin Pond argillite	TPA	*laminated argillite; minor cobble and boulder, polymictic conglomerate; rare sandstone and wacke*
Inside Gull Pond tuff	IGPT	*fine grained, buff-weathered, lithic tuff; local breccia composed of pink rhyolite fragments; welded lithic–crystal tuff*
Dinner Box Hill rhyolite	DBHR	*flow-banded and massive, pink, locally phenocrystic lava; rare rhyolite breccia*

HNPP	Hawks Nest Pond porphyry: biotite, quartz–feldspar porphyry
OPG	Otter Point granite: biotite, potassium feldspar-megacrystic granite
LPG	La Poile granite: biotite, potassium feldspar-megacrystic granite

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LEGEND

(Concluded from previous page)

Northwest Pond Brook grit

NPBG *poorly sorted, grey-brown, massive and stratified, locally graded and cross-bedded, quartz–feldspar grit and tuffaceous wacke; rip-up clasts and large blocks of laminated argillite; clasts of metamorphic and plutonic rocks*

Cinq Cerf Brook conglomerate

CCBC *thick-bedded, polymictic, boulder and cobble conglomerate interstratified with tuffaceous sandstone; local and exotic clasts of volcanic, sedimentary, plutonic and metamorphic rocks*

Grand Bruit Gull Pond tuff

GBGPT *fine grained, light green, lithic tuff; red and purple lithic tuff and slate; quartz-pebble conglomerate*

ORDOVICIAN

Bay du Nord Group

BDNG unseparated metavolcanic and metasedimentary schist

CAMBRIAN

EPG Ernie Pond gabbro: hornblende-bearing gabbro and diorite; minor pyroxenite

RG$_2$ Roti Granite: medium grained, blue-quartz-bearing tonalite and porphyry

WHG Western Head granite: biotite–hornblende granodiorite

RG$_1$ Roti Granite: coarse grained, blue-quartz-bearing granodiorite

PRECAMBRIAN–CAMBRIAN

phyllite units

Roti Point conglomerate

RPC *red-coloured, polymictic, boulder and cobble conglomerate interstratified with thin-bedded tuff; clasts of vein quartz, jasper, tuff and foliated granitoids*

Woody Duck Island tuff

WDIT *fine grained, light green, lithic tuff; red slate and tuffaceous sandstone*

Round Pond tuff

RPT *pink, quartz (± feldspar) crystal tuff*

Third Pond tuff

TPT *green lithic tuff; minor green agglomerate; rare beds of laminated argillite*

Whittle Hill sandstone

WHS *well-bedded, buff-weathered, quartz-rich sandstone; green and grey, laminated argillite; grey slate*

WHS$_C$ *conglomerate subfacies of the Whittle Hill sandstone; quartz-cobble conglomerate; polymictic cobble conglomerate containing clasts of foliated granitoids*

PRECAMBRIAN

schist-gneiss units

Cinq Cerf gneiss

CCG **paragneiss and schist; amphibolite gneiss**

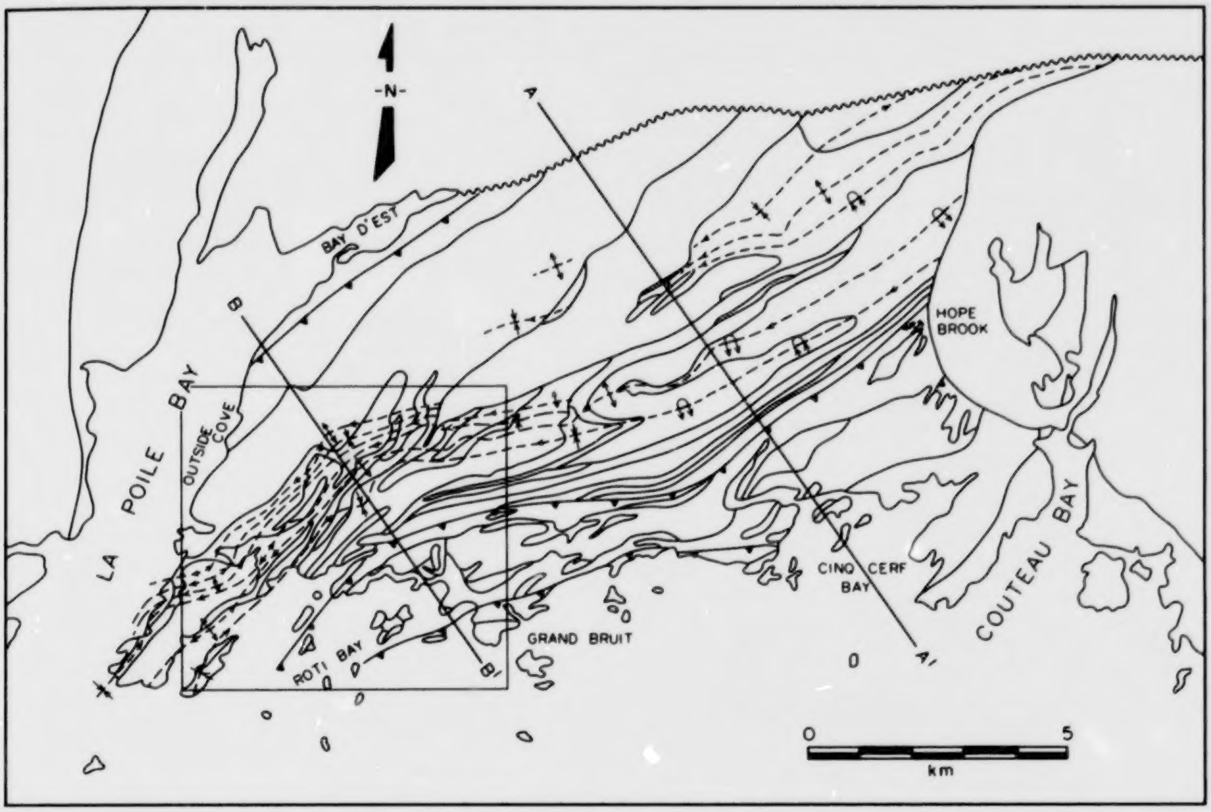


Figure 3. *Map showing the major structures of the La Poile Bay–Couteau Bay region on 1:100,000 scale. Traces of major folds and major faults are illustrated with dashed lines and solid barbed lines, respectively. Solid lines represent traces of lithological unit boundaries. See Figure 2 for legend and Figure 4 for key to symbols. Note areas where major folds are overturned and the location of the Hope Brook Gold Mine.*

of recently recognized Precambrian, Ordovician and Silurian metasedimentary and metavolcanic rocks (Dunning *et al.*, 1988) is controlled by three major faults; namely, the Grand Bruit (GBFZ), Bay d'Est (BDFZ) and Cinq Cerf (CCFZ) fault zones (O'Brien, 1987, 1988). The Bay d'Est and Cinq Cerf mylonite zones formed during Silurian D_1 shortening as synfolding thrusts (fold slides), defining the bottom and top, respectively, of a 5-km-thick thrust sheet (Figures 2 and 3). Greenschist facies rocks within the thrust sheet represent all of what is locally preserved of the recently dated Silurian La Poile Group. Lower–Middle Ordovician rocks regionally metamorphosed to the amphibolite facies between the Early Ordovician and Late Silurian (O'Brien *et al.*, 1986), occur beneath the Bay d'Est thrust and, in contrast to this writer's previous correlations, do not recur southeast of the Cinq Cerf or Grand Bruit fault zones. The Grand Bruit fault zone separates greenschist facies and upper amphibolite facies metamorphic rocks (Chorlton, 1978; O'Brien, 1988). The earliest displacements in the Grand Bruit fault zone were probably related to pre-Silurian folding and thrusting (see below), although the extent and nature of syn- to post-Silurian movements in the fault zone are unknown.

REVISION OF STRATIGRAPHICAL SUCCESSION AND STRATIGRAPHICAL NOMENCLATURE

Recent geological and geochronological studies have produced significant information about the La Poile Bay–Couteau Bay region (Stewart, 1988a,b; Yule, 1987; Stewart and Stewart, 1988; Vaskovic and Wilton, 1988; Dunning *et al.*, 1988; Dunning *et al.*, *in preparation;* Dunning and O'Brien, *in press;* G. Dunning, personal communication, 1988). Refinements to the preliminary map of O'Brien (1987) and the legend (O'Brien, 1988) reflect the expanded data base and these changes have been incorporated into the 1:100,000 scale compilation map and legend presented here (Figure 2).

The most important revisions of the stratigraphical succession and nomenclature of the La Poile Bay–Couteau Bay region are listed below. The reader should consult a Table of Formations (O'Brien, 1987, Table 1, page 313) for a review of previous age and nomenclatural assignments and for the description of numbered units referred to in the following paragraphs.

1) Amphibolitic gneiss and schist (Unit 1 of O'Brien, 1987), subsequently referred to as the Cinq Cerf gneiss (CCG of O'Brien, 1988), already possessed gneissose and schistose foliation prior to intrusion of the latest Precambrian–earliest Cambrian (Dunning *et al.*, 1988, Dunning and O'Brien, *in press)* phase of RG_1 Roti Granite. CCG is now considered a Precambrian metamorphic unit.

2) Metasedimentary schist (Unit 2 of O'Brien, 1987) was correlated with the Bay du Nord Group (O'Brien, 1988) on the basis of lithological similarity, structural style and metamorphic grade, relative to the Silurian La Poile Group. However, it contained amphibolitic interlayers and possessed schistosity prior to the intrusion of granodiorite grouped with the RG_1 phase of the Roti Granite. It is, therefore, reassigned to the Precambrian Cinq Cerf gneiss.

3) Unseparated metasedimentary and metavolcanic formations of the Bay du Nord Group (Unit 3 of O'Brien, 1987; BDNG of O'Brien, 1988) only occur northwest of the Bay d'Est fault zone. Peak metamorphism of the Lower–Middle Ordovician (Chorlton and Dallmeyer, 1986) Bay du Nord Group probably occurred prior to the earliest deformation and metamorphism of the La Poile Group (O'Brien, 1988), although both groups constitute Lower Paleozoic metamorphic units. This may corroborate suggestions that the early metamorphism of the Bay du Nord Group was synchronous with eruption of La Poile Group volcanic rocks (Dunning *et al.*, *in preparation)*.

4) Siliciclastic sedimentary rocks (Unit 6 of O'Brien, 1987) have been separated into three lithostratigraphical units (O'Brien, 1988). The Outside Gull Pond Slate (OGPS) and the Georges Pond Slate (GPS) comprise part of the Georges Brook Formation of the Silurian La Poile Group. The Whittle Hill sandstone (WHS) has been reassigned to an older, unnamed group of metasedimentary and metavolcanic rocks (see following sections on the Grand Bruit fault zone and the Cinq Cerf hanging wall). The RG_1 Roti Granite provides the minimum depositional age of the Precambrian Whittle Hill sandstone, whereas the RG_2 Roti Granite delimits its pre-Ordovician metamorphic age.

5) In the Grand Bruit–Cinq Cerf area, felsic tuff and quartz pebble conglomerate, formerly included with Unit 6 of (O'Brien, 1987), are now separated from it and grouped together to form the Grand Bruit Gull Pond tuff of (O'Brien, 1988). GBGPT is now considered to be the oldest member of the La Poile Group and is thought to stratigraphically underlie Unit 7 polymictic boulder conglomerate (O'Brien, 1987), which is herein renamed the Cinq Cerf Brook Conglomerate (CCBC). Units 8, 9, 11 and 12 of O'Brien (1987) have been called the Dinner Box Hill rhyolite (DBHR), Northwest Pond Brook grit (NPBG), Cross Gulch tuff (CGT) and Phillips Brook rhyolite (PBR), respectively (O'Brien, 1988). Unit 10 is now separated into the Twin Pond argillite (TPA), Inside Gull Pond tuff (IGPT) and Gallyboy Harbour tuff (GHT). The revised outcrop pattern of DBHR (Unit 8) shows it to be stratigraphically pinched out between NBPG and IGPT but also to be regionally infolded with these units (Figure 2).

6) In the La Poile Bay–Roti Bay area, felsic lithic tuff and overlying polymictic boulder conglomerate, originally correlated with GBGPT and CCBC (O'Brien, 1988), are herein renamed the Woody Duck Island tuff (WDIT) and the Roti Point conglomerate (RPC), respectively. Intruded by the Ernie Pond gabbro (EPG) and the RG_2 Roti Granite, the WDIT and RPC units have been separated from the Silurian La Poile Group. The Woody Duck Island tuff appears to be in conformable contact with the Round Pond tuff (RPT) and the presumably Precambrian Whittle Hill sandstone. In the La Poile Bay–Couteau Bay region, mafic tuff and agglomerate are restricted to Unit 5 of O'Brien (1987), later named the Third Pond tuff (TPT) by O'Brien (1988). These mafic volcanic rocks are interstratified with cross-bedded quartz wacke and argillite of the Whittle Hill sandstone and quartz–feldspar crystal tuff belonging to the Round Pond tuff. Rare felsic volcanic layers in WHS wacke and local bombs of felsic volcanic material in TPT agglomerate indicate that mafic and felsic volcanism occurred together with sedimentation during deposition of this unnamed group of Precambrian–Cambrian metasedimentary and metavolcanic rocks.

7) A previously unseparated and unnamed unit of felsic tuff and less common sandstone comprising the uppermost part of the La Poile Group in the La Poile Bay–Roti Bay area (O'Brien, 1988) is here correlated with the Cross Gulch tuff (CGT), which outcrops in the Grand Bruit–Cinq Cerf area (O'Brien, 1987).

CINQ CERF FAULT ZONE (CCFZ)

Fault rocks of the Cinq Cerf zone everywhere separate the Silurian La Poile Group from an unnamed group of Precambrian–Cambrian phyllite units (Figure 2; Dunning and O'Brien, *in press)*. In the Grand Bruit–Cinq Cerf area, the fault zone is marked by moderately southeast-dipping mylonite or by platy or straightened rocks showing large total strains. In the La Poile Bay–Roti Bay area, a steeply inclined zone of cataclasite is developed along part of the southeasternmost body of Silurian Hawks Nest Pond porphyry and is host rock to a northeast-trending, microspherulitic dyke of the Chetwynd Granite. Near the head of Little Roti Bay, the east-northeast-trending cataclasite zone transects a northeast-trending, subvertical, Silurian D_1 platey zone at the base of the Outside Gull Pond Slate (lower-middle La Poile Group).

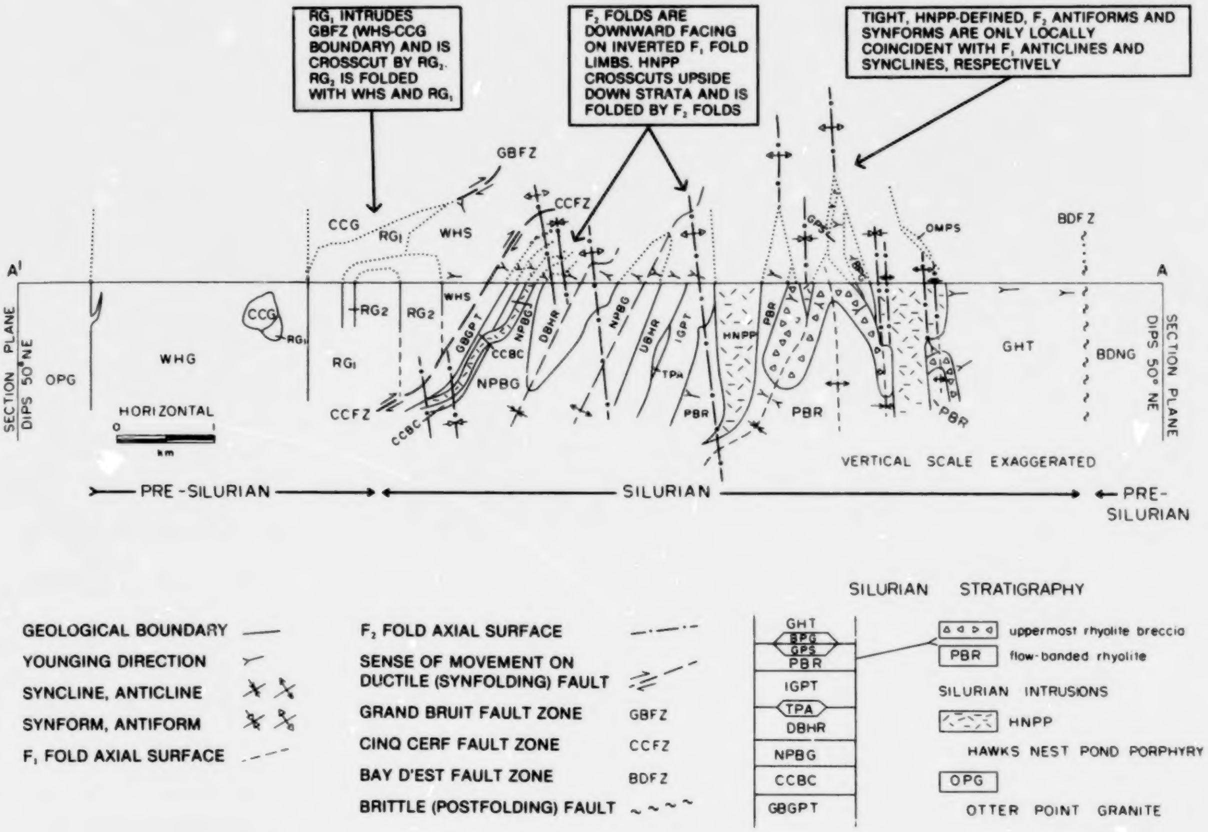

RG₁ INTRUDES GBFZ (WHS-CCG BOUNDARY) AND IS CROSSCUT BY RG₂. RG₂ IS FOLDED WITH WHS AND RG₁
F₂ FOLDS ARE DOWNWARD FACING ON INVERTED F₁ FOLD LIMBS. HNPP CROSSCUTS UPSIDE DOWN STRATA AND IS FOLDED BY F₂ FOLDS
TIGHT, HNPP-DEFINED, F₂ ANTIFORMS AND SYNFORMS ARE ONLY LOCALLY COINCIDENT WITH F₁ ANTICLINES AND SYNCLINES, RESPECTIVELY
GBFZ
CCFZ
CCG
RG₁
WHS
BDF Z
OMPS
GPS
A'
A
SECTION PLANE DIPS 50° NE
OPG
WHG
CCG
RG₁
RG2
RG2
RG₁
WHS
GBGPT
NPBG
CCBC
NPBG
DBHR
NPBG
DBHR
IGPT
TPA
HNPP
PBR
PBR
PBR
PBR
GHT
BDNG
CCFZ
CCBC
HORIZONTAL
0
1
km
VERTICAL SCALE EXAGGERATED
PRE-SILURIAN
SILURIAN
PRE-SILURIAN
GEOLOGICAL BOUNDARY
YOUNGING DIRECTION
SYNCLINE, ANTICLINE
SYNFORM, ANTIFORM
F₁ FOLD AXIAL SURFACE
F₂ FOLD AXIAL SURFACE
SENSE OF MOVEMENT ON DUCTILE (SYNFOLDING) FAULT
GRAND BRUIT FAULT ZONE GBFZ
CINQ CERF FAULT ZONE CCFZ
BAY D'EST FAULT ZONE BDFZ
BRITTLE (POSTFOLDING) FAULT
SILURIAN STRATIGRAPHY
GHT
BPG
GPS
PBR
IGPT
TPA
DBHR
NPBG
CCBC
GBGPT
uppermost rhyolite breccia
PBR flow-banded rhyolite
SILURIAN INTRUSIONS
HNPP
HAWKS NEST POND PORPHYRY
OPG
OTTER POINT GRANITE

Figure 4. *Representative geological section of the Grand Bruit–Cinq Cerf area. Horizontal scale is 1:50,000, vertical scale is exaggerated and section plane dips 50° NE. Insets summarize and identify important geological relationships. Folds are not necessarily seen in profile. See Figure 3 for line of section.*

CCFZ Footwall

In section AA' (Figures 3 and 4), the Cinq Cerf fault zone is a Silurian D_1 thrust zone, across which a hanging wall sequence of Precambrian–Cambrian stratified rocks and pre-Silurian intrusive rocks are translated northwestwards, over a footwall sequence of Silurian stratified and intrusive rocks. Beneath the synfolding thrust or fold slide, the La Poile Group is isoclinally folded by southeasterly inclined, overturned, major F_1 folds that display consistent S-shapes and sinistral vergence (Figures 3 and 4). In the Grand Bruit–Cinq Cerf area, for at least 3 km beneath the D_1 thrust, inverted members of the La Poile Group predominate over right-way-up members. This is consistent with common observations of upward-facing, S_1 slaty cleavage, lying more gently than bedding in upside-down Silurian phyllites (O'Brien, 1987).

The gentle to moderate plunge of the S-shaped, F_1 major folds in the Grand Bruit–Cinq Cerf area means that the stratigraphical order of La Poile Group members can be discerned from a map by looking southwest down-the-plunge of fold hinge zones (Figure 3) or from a section by looking northwest on the long limbs of regional folds (Figure 4). By doing so, it is evident that the Cinq Cerf fault zone is regionally transgressive relative to La Poile Group stratigraphy, the amount of stratigraphical separation apparently increasing southwestward. Therefore, as the thrust zone is traced northeastward, progressively older parts of the La Poile Group are presumably brought from subcrop to outcrop. The oldest part of the Grand Bruit Gull Pond tuff (lowermost La Poile Group) occurs near the Hope Brook Gold Mine where, regionally inverted and unaltered, it is intruded by the Silurian Hawks Nest Pond porphyry and is terminated by the Devonian Chetwynd Granite (see also Stewart and Stewart, 1988; Stewart, 1988a).

A D_1 transition from near upright to subrecumbent structures occurs on various scales and is a characteristic feature of the La Poile Group near the D_1 mylonite belts that form its boundaries. The overturned, southeasterly inclined, major F_1 folds present in the La Poile Group beneath the southeast-dipping D_1 thrust, are paired with upright, major F_1 folds that plunge gently in the central part of the thrust

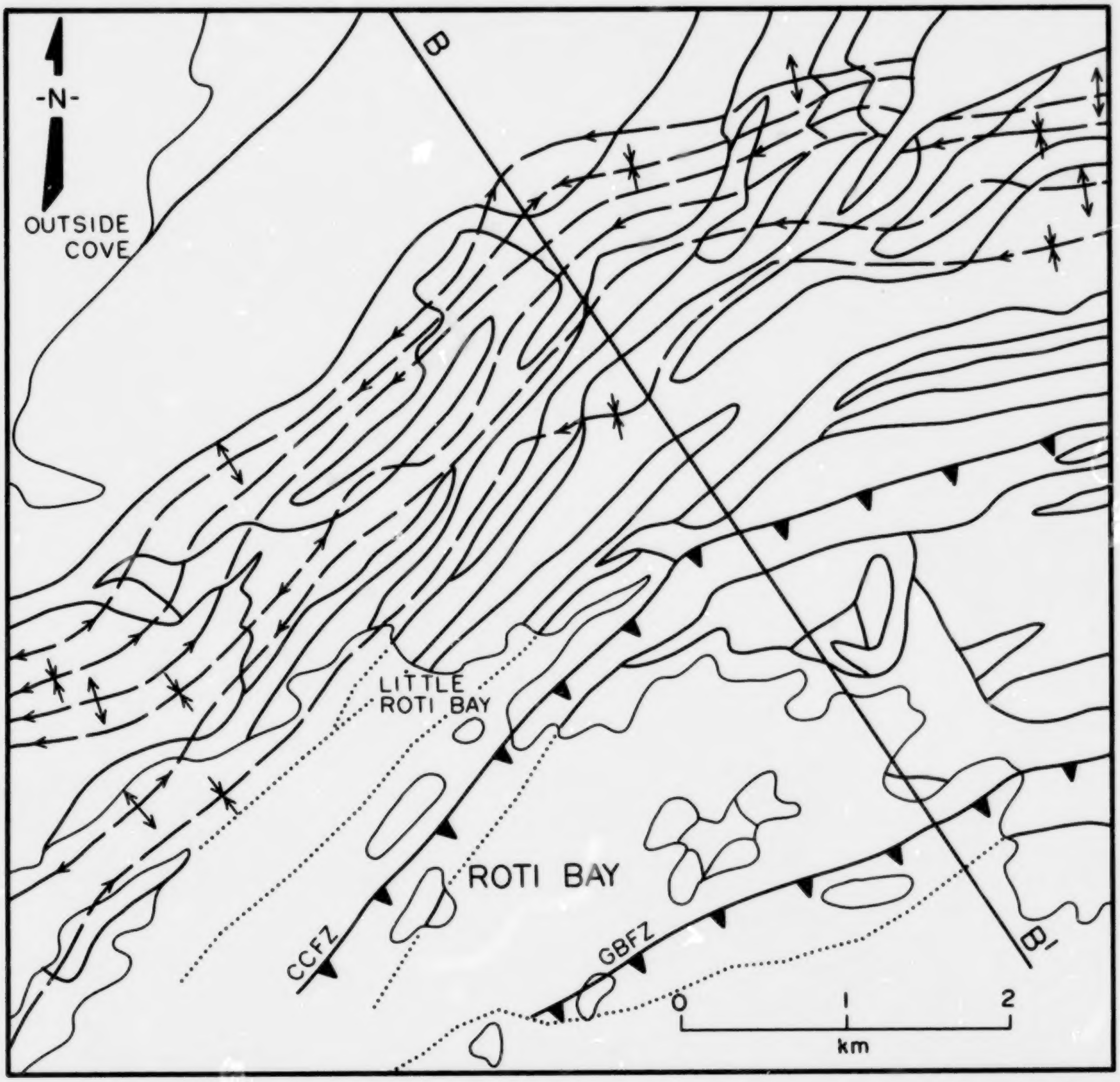
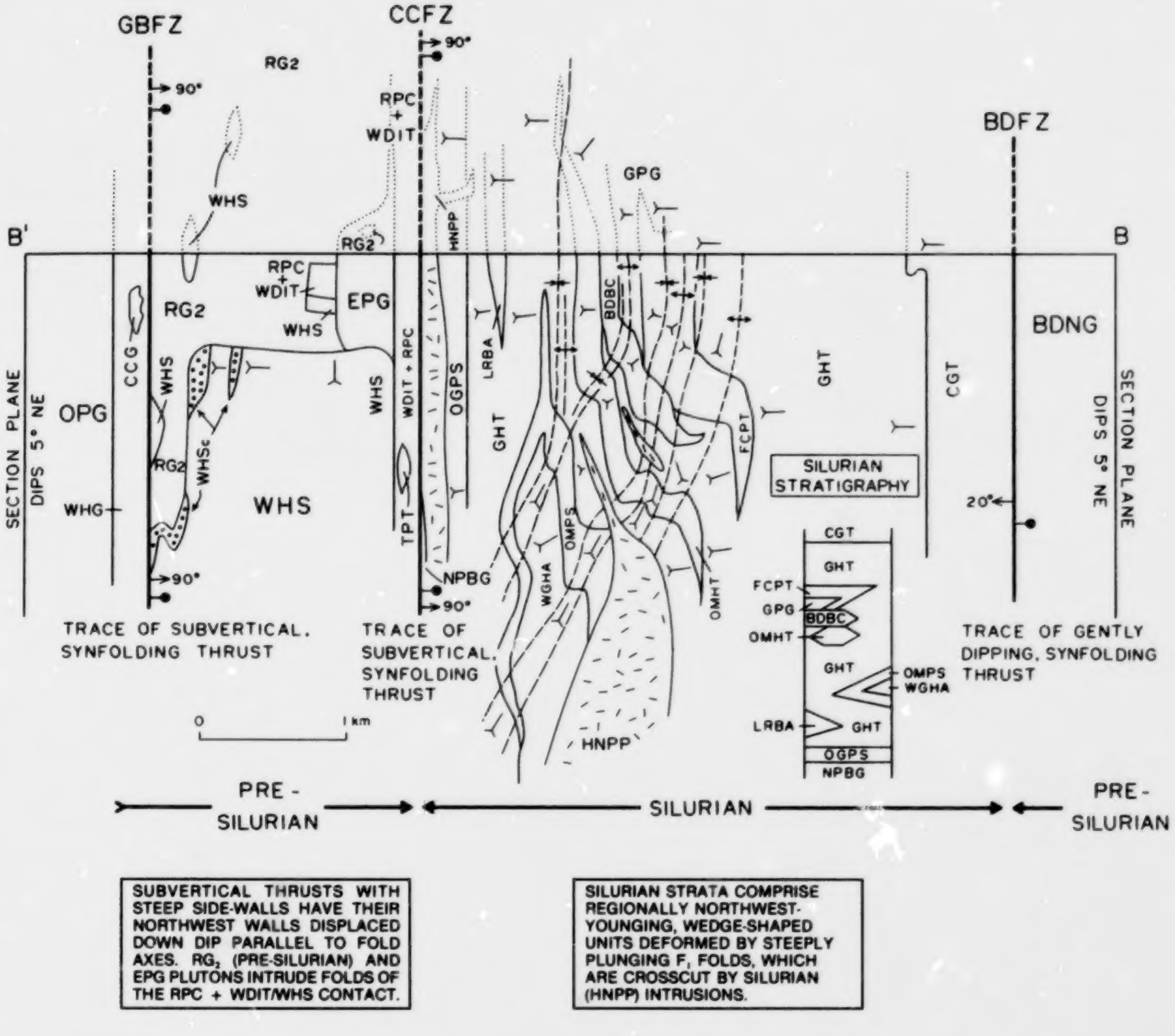


Figure 5. *Detailed map of the Roti Bay–Outside Cove area illustrating, in particular, the nature of the periclinal folds affecting members of the Georges Brook Formation of the La Poile Group northwest of the Cinq Cerf fault zone. See inset in Figure 3 for regional setting.*

sheet (e.g., the PBR-cored anticlines in section AA', Figure 4). Likewise, moderately plunging, overturned F_1 folds are traceable southwestward down-the-plunge into upright periclinal folds (Figures 2, 3 and 5). Steeply plunging anticlines–synclines, and neutral F_1 folds of thin, wedge-shaped members of the La Poile Group occur immediately northwest of the narrow 'straightened' zone at the base of OGPS in the Cinq Cerf fault zone (Figure 6). Platey OGPS rocks displaying steeply dipping S_1 foliation and vertically plunging L_1 lineation parallel to F_1 fold hinges, indicate vertical extension (translation) of the La Poile Group near an upright, D_1 fold slide. Some La Poile Group members comprising the sidewall sequence northwest of the upright fold slide, can be traced continuously into the footwall sequence beneath the gently inclined fold slide in the Grand Bruit–Cinq Cerf area. Therefore, the major D_1 structure is assumed to be a ductile listric thrust composed of vertical- and southeast-dipping segments. Regardless of the attitude of ductile fault rocks, Silurian strata moved downward relative to pre-Silurian strata everywhere along the CCFZ during D_1 deformation (Figures 4 and 6). The listric thrust forms both vertical and southeast-dipping footwall ramps in the La Poile Group; fold-related contraction faulting did not produce thrust flats.

Within the Silurian thrust plate, sheet intrusion of dykes and sills of Silurian diabase and Silurian quartz–feldspar porphyry was controlled by pre-existing F_1 folds and D_1 faults


GBFZ
CCFZ
BDFZ
90°
90°
90°
90°
20°
B'
B
RG2
RPC
+
WDIT
WHS
RG2
GPG
HNPP
RPC
+
WDIT
EPG
WHS
BDNG
BDBC
CCG
WHS
WHS
WDIT + RPC
OGPS
LRBA
GHT
GHT
CGT
SECTION PLANE
DIPS 5° NE
OPG
WHSc
RG2
FCPT
WHS
WHG
TPT
OMPS
WGHA
OMHT
NPBG
SILURIAN STRATIGRAPHY
CGT
GHT
FCPT
GPG
BDBC
OMHT
GHT
OMPS
WGHA
LRBA
GHT
OGPS
NPBG
TRACE OF SUBVERTICAL, SYNFOLDING THRUST
TRACE OF SUBVERTICAL, SYNFOLDING THRUST
TRACE OF GENTLY DIPPING, SYNFOLDING THRUST
0
1 km
HNPP
PRE-SILURIAN
SILURIAN
PRE-SILURIAN

SUBVERTICAL THRUSTS WITH STEEP SIDE-WALLS HAVE THEIR NORTHWEST WALLS DISPLACED DOWN DIP PARALLEL TO FOLD AXES. RG_2 (PRE-SILURIAN) AND EPG PLUTONS INTRUDE FOLDS OF THE RPC + WDIT/WHS CONTACT.

SILURIAN STRATA COMPRISE REGIONALLY NORTHWEST-YOUNGING, WEDGE-SHAPED UNITS DEFORMED BY STEEPLY PLUNGING F_1 FOLDS, WHICH ARE CROSSCUT BY SILURIAN (HNPP) INTRUSIONS.

Figure 6. *Representative geological section of the La Poile Bay–Roti Bay area. Horizontal scale is 1:50,000, vertical scale is slightly exaggerated and section plane dips 5° NE. Insets summarize relationships of pre-Silurian rocks between the Grand Bruit fault zone (GBFZ) and the Cinq Cerf fault zone (CCFZ) and relationships of Silurian rocks northwest of CCFZ. Folds are not necessarily seen in profile. See Figure 3 for line of section and Figure 4 for key to symbols.*

in the La Poile Group. As a result, the intrusive rocks also reflect the upright-recumbent structural transition. For example, subvertical, discordant, diabase intrusions occupy the hinge zones of EPR–GPG–FCPT-cored, upright F_1 periclines (Figures 3 and 5) in the centre of the thrust sheet; whereas, subhorizontal, concordant, diabase intrusions are emplaced into thrust-generated D_1 mylonite at the structural base of the La Poile Group.

The southeasternmost body of Hawks Nest Pond porphyry (Figures 2 and 3) intrudes several members of the lower part of the La Poile Group as the intrusion follows the course of the Cinq Cerf fault zone. In the Grand Bruit–Cinq Cerf area, the southeast-dipping, slightly transgressive, porphyry sill intrudes upside down, highly deformed strata on the inverted limb of an overturned, DBHR-cored syncline (Figure 4). Traced continuously southwestward into the La Poile Bay–Roti Bay area, the HNPP sill becomes vertical where the F_1 syncline becomes upright. Here, the porphyry is hosted by right-way-up, northwest-dipping or vertical, northwest-younging strata on the upright limb of the fold (section BB', Figure 6). In both areas, the porphyry intrusion is emplaced parallel to axial planar S_1 slaty cleavage and contains xenoliths of this foliation (O'Brien, 1988). Extensive, northeast-trending bodies of HNPP cut across the limbs and hinges of gently and steeply plunging, F_1 major folds and, thereby, transgress most members of the La Poile Group (Figures 3, 4 and 6).

A major effect on the geometry of the thrust plate caused by the emplacement of Silurian sheet intrusions is the distortion and local reorientation of F_1 folds and S_1 cleavage in the La Poile Group (O'Brien, 1987). A map showing the regional traces of F_1 axial surfaces illustrates that the F_1 folds near the northeastern and southeastern borders of the main body of the Hawks Nest Pond porphyry were either sharply transected by the sheet intrusion or displaced and locally reoriented near its intrusive margins (Figure 3). The D_1 structures affected by the Hawks Nest Pond porphyry were distorted from the regional D_1 trend prior to being overprinted by F_2 upright folds and F_3 cross folds. Bauer (1986) describes a similar example of pluton emplacement and multiple folding in the Archean of Minnesota. Patterns of F_1 axial traces in the southwesternmost part of the La Poile Group may indicate the presence of HNPP porphyry bodies at shallow depth.

The Silurian D_1 and Siluro–Devonian D_2 deformations are regionally coaxial, both producing northeast-trending structures in the La Poile Group. Interference of F_1 and F_2 folds and overprinting of S_1 and S_2 foliation are, however, restricted to areas where D_1 structures were suitably oriented to be affected by D_2 deformation. This configuration only existed where steep D_1 structures were reoriented or bent to lie flatter near intrusion margins or, near southeast-dipping thrusts, where D_1 structures originally formed with gentle inclinations. F_2 folds deform D_1 mylonites in the Cinq Cerf (Figure 4) and Bay d'Est (O'Brien, 1988) fault zones.

Quartz–feldspar porphyry and diabase intrusions of Silurian age are locally foliated by vertical S_2 cleavage and folded by steeply to gently plunging F_2 folds. In the central part of the Silurian thrust sheet, where all structures are generally steeply dipping, country-rock defined, open F_1 anticlines and synclines are locally coincident with intrusion-defined, tight F_2 antiforms and synforms, respectively (section AA', Figure 4). However, periclinal F_2 folds of the country rocks are more commonly developed on F_1 fold limbs, where they are variably upward or downward facing (e.g., NPBG-cored refolded fold in Figure 3 and in section AA' in Figure 4). Typically open and upright to asymmetrical, F_2 folds consistently form Z-shaped fold pairs that display dextral vergence. In La Poile Group phyllite, F_2 folds have axial planar S_2 crenulation cleavage.

Gold and copper mineralization is found in the central part of the Silurian thrust sheet. Pyrite and chalcopyrite occur in zones of sericitic and silicic alteration developed in the main and satellite bodies of the Hawks Nest Pond porphyry and in the PBR, NPBG, GHT, WGHA, OMHT and FCPT members of the La Poile Group peripheral to the intrusion. This period of Silurian gold mineralization is post-F_1 –pre-F_2 in age.

CCFZ Hanging Wall

Pre-Silurian rocks in the hanging wall and side wall of the D_1 listric thrust in the Cinq Cerf fault zone have a similar lithological composition, degree of metamorphism, intrusive history, structural association and alteration style as the Silurian rocks in the footwall of the structure. As a result, with the exception of the original surveyor (Cooper, 1954), all previous workers in the La Poile Bay–Couteau Bay region (Chorlton, 1978,1980; Swinden, 1981,1984; McKenzie, 1986; O'Brien, 1987,1988) assigned stratified rocks outcropping between the Cinq Cerf and Grand Bruit fault zones to the La Poile Group. In previous reports, this writer implied that the Cinq Cerf fault zone was an intra-La Poile Group splay off the Grand Bruit fault zone, which contained the D_1 thrust separating Silurian phyllite from pre-Silurian schist and gneiss (O'Brien, 1988). Furthermore, because the Whittle Hill sandstone (Figure 2) was taken to be the stratigraphically lowest member of the La Poile Group, the Hope Brook Gold Mine was considered to be hosted, at least in part, by Silurian deposits.

The common premise of all previous workers was that the intrusive rocks in the CCFZ hanging-wall sequence either formed part of the La Poile Group or were younger than it. Dunning and O'Brien *(in press)* report, however, a latest Cambrian–earliest Ordovician crystallization age for the RG_2 phase (Figure 2) of the Roti Granite. This suggests that the postulated correlations and age assignments of previous workers are in doubt. In the La Poile Bay–Roti Bay area, as illustrated in section BB' in Figure 6, the tonalite and porphyry (RG_2) phase of the Roti Granite intrudes the Whittle Hill sandstone (WHS) and the Woody Duck Island Tuff (WDIT). Because the conglomeratic subfacies of the Whittle Hill sandstone (WHSc), the Third Pond tuff (TPT), the Round Pond tuff (RPT) and the Roti Point conglomerate (RPC) are interstratified with one or both of WHS and WDIT (Figure 2), the depositional age of all the above units is presumed older than the RG_2 phase of the Roti Granite, dated as 499 $^{+3}_{-2}$ Ma (Dunning and O'Brien, *in press).*

In the Grand Bruit–Cinq Cerf area, the RG_1 phase of the Roti Granite intrudes the Whittle Hill sandstone and is reported to have yielded a latest Precambrian–earliest Cambrian (563 ± 4 Ma) crystallization age (Dunning *et al.*, 1988; Dunning and O'Brien, *in press).* This gives the minimum depositional age of the Whittle Hill sandstone as pre-earliest Cambrian. In this area, the RG_2 phase of the Roti Granite is emplaced as sheets along steeply dipping beds of the Whittle Hill sandstone. RG_2 screens cross-cut the RG_1–WHS intrusive contact (Figure 4) and are folded along with bedding-parallel phyllitic foliation in the Whittle Hill sandstone (O'Brien, 1987). This writer is uncertain if the primary foliation in the Whittle Hill sandstone is Precambrian or Cambrian in age.

The Silurian La Poile Group contains granodiorite clasts with a pre-incorporation foliation as well as clasts of porphyry injected by mafic intrusions. These fragments are lithologically similar to pre-Silurian intrusions hosted by the unnamed group of Precambrian–Cambrian sedimentary and volcanic rocks. This observation is in agreement with recently determined ages of the RG_1 and RG_2 phases of the Roti Granite (Dunning and O'Brien, *in press*). Furthermore, it indicates that deposition, intrusion, low-grade metamorphism and initial deformation of the CCFZ hanging-wall sequence took place prior to the time period in which the La Poile Group was deposited. Cooper (1954) originally alluded to an unconformity at the base of the La Poile Group, as he defined it, although he did not mention an exposed surface of unconformity. In the Grand Bruit–Cinq Cerf area, he drew the lower boundary of the group near CCBC and GBGPT; whereas, in the La Poile Bay–Roti Bay area, it was drawn near RPC and WDIT. As first suggested by Marten (1984) and re-emphasized by Stewart (1986), the Cinq Cerf shear belt or mylonite zone has probably displaced a sub-La Poile Group unconformity, which originally lay in the La Poile Bay–Couteau Bay region near what is now the site of the Cinq Cerf fault zone.

BAY D'EST FAULT ZONE (BDFZ)

Fault rocks of the Bay d'Est zone everywhere separate the Ordovician Bay du Nord Group from the Silurian La Poile Group (Figure 2; O'Brien *et al.*, 1986; Dunning *et al.*, 1988). A zone of greenschist facies mylonite, presumably derived from both groups, occurs along the southeast coast of Bay d'Est. There, the mylonite zone dips gently beneath a Silurian hanging-wall sequence of subrecumbent F_1 folded phyllites that belong to a locally inverted part of the Cross Gulch tuff (uppermost La Poile Group). It is unknown if the Ordovician footwall sequence of southeast-dipping Bay du Nord schist is inverted or right-way-up beneath the Bay d'Est mylonite zone. However, in contrast to the Cinq Cerf fault zone, younger rocks overthrusted older rocks during Silurian D_1 movements in the Bay d'Est fault zone.

Where BDFZ mylonites dip subvertically in the fault zone, Bay du Nord Group rocks in the sidewall display an upright, bedding-parallel, S_2 crenulation cleavage that is associated with transposition and grain-size reduction of S_1 schistosity and retrogression of amphibolite facies schist (O'Brien, 1987,1988). This crenulation cleavage is also regionally expressed in the southeast-dipping footwall sequence well below the thrust. Here, it occurs as a sporadically developed, locally downward facing, spaced foliation that is axial planar to gently plunging, upright folds. These secondary folds refold southeasterly inclined, peak-metamorphic, intrastratal folds in the Bay du Nord Group. A similarly oriented, commonly downward facing, crenulation cleavage in hanging wall phyllite, developed during Siluro–Devonian D_2 deformation (O'Brien, 1987, 1988). Strong, inhomogeneous D_2 deformation of footwall ramps is a common feature of the Bay d'Est fault zone that is not observed in the Cinq Cerf structure.

The mylonite zone outcropping along the Bay d'Est coast is transected, near the head of the bay, by a fault zone of incohesive, random-fabric breccia and gouge (Figure 3). A steeply dipping structure within the Bay D'Est fault zone, it is marked by a distinct topographical depression throughout the La Poile Bay–Couteau Bay region. Unlike the D_1- and D_2-activated ductile fault, the brittle fault cross-cuts many of the major folds in the La Poile Group as well as the Devonian Chetwynd Granite (Figure 3). The matrix texture of the fault rocks, together with the presence of schist, phyllite and mylonite fragments, probably indicate that brittle faulting occurred after regional metamorphism of the Bay du Nord and La Poile groups. Quartz–chlorite–potassium feldspar veins invade gouge and breccia, and are associated with patchy silicification, sericitization, pyritization and gold mineralization.

GRAND BRUIT FAULT ZONE (GBFZ)

At some time prior to Cambro–Ordovician intrusion and Silurian deposition, the Grand Bruit fault zone probably formed the northwestern boundary of exposed Cinq Cerf gneiss. On the scale of the La Poile Bay–Couteau Bay region, the Grand Bruit fault zone separates a hanging-wall sequence of Precambrian Cinq Cerf gneiss from a footwall sequence of unnamed Precambrian–Cambrian metasedimentary and metavolcanic rocks (Figure 2). However, on more detailed scales, intrusions within the fault zone intervene between gneiss and phyllite, and are commonly mylonitized along with gneiss or phyllite. Along the coast, GBFZ mylonite dips vertically or steeply southeast. However, in the northeastern part of the region near the Hope Brook Gold Mine, subhorizontal platey zones occur near the base of the Cinq Cerf gneiss (and the GBFZ ?).

The Cinq Cerf gneiss is narrowest near Roti Bay and widest north of Couteau Bay (Figures 2 and 3). On the Roti Bay islands, it is intruded by the Otter Point granite, which is assumed to terminate the CCG outcrop belt southwestward (Figure 2). In the northeastern part of the region, the Cinq Cerf gneiss is intruded by the posttectonic Chetwynd Granite, although a large roof pendant within the intrusion contains the on-strike extension of the CCG outcrop belt. Therefore, in the area surveyed, the Grand Bruit fault zone is transected by the Silurian Otter Point Granite and Devonian Chetwynd Granite (Figure 2).

In the La Poile Bay–Couteau Bay region, intrusions of known and presumed pre-Silurian age are restricted to the Grand Bruit fault zone, the GBFZ footwall and the GBFZ hanging wall. The 563 ± 4 Ma phase of the Roti Granite (RG_1) is interpreted to be emplaced along the Grand Bruit fault zone (Figure 4). Along its northwestern boundary, RG_1 Roti Granite intrudes the Whittle Hill sandstone; along its southeastern boundary, RG_1 Roti Granite contains xenoliths of Cinq Cerf gneiss. In places, only 250 m of this lenticular granite separate the Whittle Hill sandstone and the Cinq Cerf gneiss. RG_1 Roti Granite is considered a stitching pluton linking GBFZ hanging wall and GBFZ footwall because, at the time of latest Precambrian–earliest Cambrian intrusion, the CCG unit was already gneissic and yet lay very close to the WHS unit, which was never regionally metamorphosed above chlorite-grade. The overall distinction between WHS and CCG units, makes it unlikely that the pre-RG_1 regional metamorphic gradient was notably steep and that the Whittle Hill sandstone comprised part of a succession that was the protolith of the Cinq Cerf gneiss (Cooper, 1954; Chorlton, 1978,1980). Rather, their pre-RG$_1$ relationship was probably faulted or unconformable; this writer favours the former interpretation because of the emplacement mode of RG$_1$ Roti Granite (O'Brien, 1988).

The GBFZ footwall is host to a variety of pre-Silurian intrusions, which extend along the strike of the footwall for considerable distances but do not occur in the GBFZ hanging wall. The same relationship applies to presumed pre-Silurian intrusions that occur exclusively in the hanging wall. Most of these variably deformed intrusive units are younger than the RG$_1$ pluton and, therefore, were emplaced after the presumed earliest displacements in the Grand Bruit fault zone. Nevertheless, continued movements along the structure apparently partitioned the intrusions so that they presently comprise either footwall or hanging-wall intrusive suites.

The GBFZ footwall is host to the RG$_2$ Roti Granite, the Ernie Pond gabbro and intrusive felsite bodies; the latter is not shown on the legend of Figure 2. The Western Head granite is spatially associated with the Cinq Cerf gneiss in the GBFZ hanging wall. Mafic dyke swarms and sheets of quartz-feldspar porphyry are present throughout the footwall and the hanging wall; however, they are most intensely deformed in the Grand Bruit mylonite zone.

Comparison of sections drawn through the GBFZ hanging wall and footwall gives an indication of the protracted history of intrusion and differential displacement that occurred near the Grand Bruit fault zone (Figures 4 and 6). In section BB', the contact of the Cambro–Ordovician RG$_2$ Roti Granite dips vertically and cross-cuts vertical, northwest-younging, WHSc conglomerate and southwest-younging, WHS sandstone on the limb and in the hinge zone, respectively, of northeast-trending, steeply southwest-plunging, major folds. The Woody Duck Island tuff (WDIT) and the Roti Point conglomerate (RPC) are also deformed by these folds, which have vertical, axial planar, slaty cleavage. Although only the Whittle Hill sandstone occurs as variably deformed, map-scale enclaves within RG$_2$ Roti Granite, the Woody Duck Island tuff, Roti Point conglomerate and Whittle Hill sandstone were regionally folded so that all units had a northwest strike prior to being intruded by the RG$_2$ pluton and the younger EPG pluton. Contact metamorphic porphyroblasts in hornfelses around bosses of Ernie Pond gabbro overgrow the regional slaty cleavage in several of the Precambrian–Cambrian phyllite units. Porphyroblasts are locally distorted, however, especially where the margins of the Ernie Pond gabbro are deformed. The intrusive contact between the RG$_2$ and WHS units is itself folded by map-scale vertical folds, which have highly variable apical angles dependent upon their position in the hanging wall relative to the main fault zone (Figures 2 and 6).

With increasing proximity to the Grand Bruit fault zone, stratified rocks in the footwall sequence display relatively more folds that display axial planar slaty cleavage, although on continuously decreasing orders of magnitude. In general, as the number of minor folds increases, so does the amount of minor sheet intrusions in the Whittle Hill sandstone. Regardless of size, all of these folds tighten and pass into upright, platey zones in which mafic sills are invariably foliated. Within the fault zone proper, the Whittle Hill sandstone and the RG$_2$ Roti Granite are strongly deformed and sheared together with mafic sills in upright zones of mylonite characterized by vertical extension lineations. Here, crenulation cleavage is locally developed in mafic sills that, along with previously mylonitized host rocks, define secondary, highly curvilinear, isoclinal folds.

Where the Grand Bruit fault zone forms the boundary of footwall units as young as the Cambro–Ordovician RG$_2$ Roti Granite (Figures 2 and 6), the hanging wall is locally composed of undated Western Head granite that contains a peripheral belt of mylonite and highly sheared mafic sills. In parts of the GBFZ hanging wall, away from the fault zone, undeformed Western Head granite intrudes foliated RG$_1$ Roti Granite and Cinq Cerf gneiss, and contains enclaves of both of these units. The intrusive contact between the Eocambrian RG$_1$ Roti Granite and the Precambrian Cinq Cerf gneiss is locally obliterated within the mylonite belt that juxtaposes the Western Head granite against several Ordovician and older footwall units. Fault movement along this particular structure in the Grand Bruit fault zone is likely to be either Ordovician or Silurian in age. Sheared mafic dykes in the deformed margin of the Otter Point granite attest to the Siluro–Devonian deformation near its intrusive contact with the Western Head granite and the GBFZ hanging wall.

In section AA' (Figure 4), an early GBFZ thrust defining the original WHS–CCG boundary is assumed to be occupied by the RG$_1$ pluton. The RG$_1$ Roti Granite and concealed Precambrian thrust are terminated by the Western Head intrusive body. The structurally lowest part of the Cinq Cerf gneiss is assumed to dip gently and to lie above RG$_1$ Roti Granite and present erosion level between Cinq Cerf Bay and the Hope Brook area, where it again comes to outcrop (Figures 2 and 3). The vertical northwestern margin of the Western Head granite transects steep CCG with concordant RG$_1$ on the coast and flat CCG with concordant RG$_1$ near the Hope Brook Gold Mine. Therefore, the presumed pre-RG$_1$ thrust had acquired its listric shape prior to intrusion of the Western Head granite. Despite being present throughout the GBFZ footwall, a major vertical swarm of RG$_2$ Roti Granite screens are emplaced along parts of the upright RG$_1$–WHS contact (Figures 2 and 4). This occurred after local *C–S* fabrics developed in the RG$_1$ Roti Granite and possibly, simultaneously, with some of the folding and shearing of the Whittle Hill sandstone. It appears, therefore, that there has been a longevity of Proterozoic and Phanerozoic ductile faulting in the Grand Bruit fault zone.

In the La Poile Bay–Couteau Bay region, extensive gold mineralization is predominantly located in the GBFZ footwall, although much rarer, less significant occurrences are found in the GBFZ hanging wall. In all locations, gold mineralization is associated with silicification, sericitization and pyritization of pre-Silurian intrusions or their country rocks. Altered rocks include lithologies typical of the RG_1, RG_2, WHS and CCG units. At the Hope Brook Gold Mine, cordierite-bearing mafic dykes forming one of the regional swarms in the GBFZ footwall are affected by high alumina alteration but contain xenoliths of completely silicified rocks (see also Stewart and Stewart, 1988; Stewart, 1988b).

CONCLUSIONS

The conclusions to be drawn from this study are:

1) The region between La Poile Bay and Couteau Bay is underlain by variably deformed and metamorphosed, sedimentary and volcanic rocks that comprise four distinct belts of lithostratigraphical units separated by three major faults belonging to the Hermitage Flexure system. The Bay d'Est fault zone forms the southeastern limit of the Ordovician Bay du Nord Group, the Cinq Cerf fault zone marks the southeastern boundary of the Silurian La Poile Group, and the Grand Bruit fault zone separates an unnamed Precambrian–Cambrian group of phyllite units from the Precambrian Cinq Cerf gneiss.

2) The earliest formed structures in the Bay d'Est, Cinq Cerf and Grand Bruit fault zones are synfolding thrusts or fold slides, which were produced in association with an upright-recumbent structural transition. Precambrian, Precambrian–Cambrian, Ordovician and Silurian metasedimentary and metavolcanic rocks are separately disposed in four thrust sheets bounded by southeast dipping, listric thrusts. In the La Poile Bay–Couteau Bay region, northwestward translation of thrust sheets occurred during Silurian D_1 contraction faulting. With the exception of the Bay d'Est structure, all rocks in the hanging walls of these thrusts are older than the rocks comprising their footwalls. The earliest thrust-related structures in the Grand Bruit fault zone are, however, presumed to be older than those in the Cinq Cerf and Bay d'Est fault zones. Pre-Silurian movements in the Grand Bruit fault zone antedate the intrusion of the Western Head granite. Differential displacement preceded emplacement of Cambro–Ordovician RG_2 Roti Granite and mafic dyke swarms, and is probably related to the initial folding of the Precambrian–Cambrian metasedimentary and metavolcanic rocks.

3) Widespread deformation and low-grade regional metamorphism continued after the thrust sheets were assembled. Many of the metasedimentary and metavolcanic rocks in the La Poile Bay–Couteau Bay region were deformed to varying degrees after the intrusion of the Hawks Nest Pond porphyry, the La Poile granite and the Otter Point granite or, locally, by the emplacement of these plutons and batholiths. The Silurian–Devonian D_2 deformation was most intense near Silurian intrusions or major faults. However, it deformed most areas where Silurian D_1 or older structures were suitably oriented to be affected by D_2 deformation. A transition from upright to recumbent structures is seen in country rocks marginal to intrusions; however, in contrast to the earlier compression-related transition, this feature developed as the thrust sheets were extended to make space for resurgent Silurian magma. Subsequent 'hanging wall-down–footwall-up' movements in the Cinq Cerf and Bay d'Est fault zones may be responsible for producing the F_2 folds in the Silurian thrust sheet or, alternatively, their development may herald a return to a compressive regime.

4) In the La Poile Bay–Couteau Bay region, intrusions of Precambrian, Cambrian, Ordovician, Silurian and Devonian age, serve to bracket distinct intervals of deposition, deformation, metamorphism and alteration. Some events are unique to particular thrust sheets; others have affected most rocks in the region. Precambrian, Ordovician and Silurian geological events are significantly distinct. However, because the same types of processes were operative in the region from Precambrian to Devonian, rocks formed during this period are generally of similar aspect.

5) Gold mineralization occurs throughout the La Poile Bay–Couteau Bay region and is hosted, in order of increasing abundance, by rocks of the Precambrian Cinq Cerf gneiss, the Ordovician Bay du Nord Group, the Silurian La Poile Group and an unnamed group of Precambrian–Cambrian phyllite units. Sericitic, pyritic and silicious alteration is commonly associated with gold-bearing felsic intrusions within the thrust sheets or within the major fault zones. Gold was apparently introduced or remobilized at several times in different areas. Mineralization occurred sporadically as the fault zones evolved from early stage, synmetamorphic overthrusts affecting basement and cover to late stage, brittle faults. The latter structures are developed in the youngest unit in the region, the epizonal Chetwynd Granite (Dickson *et al., this volume)* and probably formed at shallow levels below a Devono–Carboniferous paleosurface similar to that described by Chorlton (1983).

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CONODONT FAUNAS FROM THE CATCHERS POND AND CUTWELL GROUPS, CENTRAL NEWFOUNDLAND

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ABSTRACT

Conodonts have been extracted from carbonates of the Catchers Pond and Cutwell groups, of Central Newfoundland. The Catchers Pond Group conodonts are found to range through a longer time period than previously reported, and are from early to middle Arenig in age.

Fossils have been reported previously from the Cutwell Group of late Arenig–early Llanvirn age. The new data indicate that the age of the conodonts is late Arenig to Llanvirn, and even possibly as young as basal Llandeilo.

The similar Colour Alteration Index (CAI), conodont assemblage and zircon U–Pb age from the Cutwell Group suggest that it is a correlative of the Buchans Group.

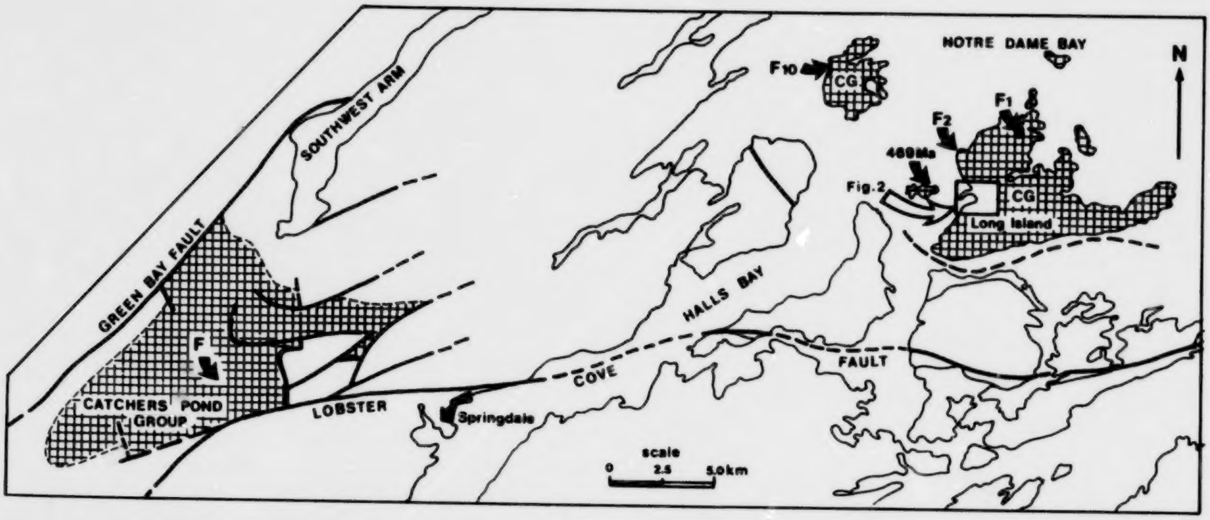

Figure 1. *Distribution of the Catchers Pond Group and the Cutwell Group in central Newfoundland. CG–Cutwell Group, Long Island and Little Bay Island; F–conodont sample localities.*

INTRODUCTION

Three samples from a limestone bed of the Catchers Pond Group were processed for conodonts, in order to confirm the early Arenig age of the fauna reported by Bergström *et al.* (1974) and to establish the facing direction of the bed.

The Cutwell Group (Figure 1) is structurally complex and much of the section is considered to be repeated thrusting (Szybinski, 1988, *this volume*). Fifty samples from various carbonate lithologies were collected in order to correlate the different thrust sheets. At the same time, an attempt was made to compare the fauna from these samples with those reported by Meyer *et al.* (1988) and Stouge (1980).

The extensive fauna that was retrieved from both groups, together with the large, and as yet only partially described macrofauna (Strong and Kean, 1972), not only made correlation between thrust slices possible, but also enabled the environment of deposition to be reconstructed.

THE CONODONT FAUNAS

Catchers Pond Group

Three large samples of highly recrystalized and altered limestone from the east–west-trending bed in the middle of this mainly volcanic group (Figure 1) were processed for conodonts. The extracted conodonts were extremely small in size, well preserved and with a very high Colour Alteration Index (CAI) of 6.5 to 7. The good preservation of the conodonts and their high CAI suggest that the alteration of this limestone bed is mostly due to an increase in temperature to at least 490° C (Rejebian *et al.*, 1987).

Two samples from the southern part of the limestone bed contain an assemblage dominated by *Paracordylodus gracilis* Lindström, *Prioniodus cf. P. elegans* Pander, and simple cones. The other sample from the northern part of the bed contained not only a similar simple cone fauna and *Prioniodus cf. P. elegans* Pander, but also *Oepikodus evae* Lindström, indicating that the age of the fauna ranges from the early Arenig *Prioniodus elegans* Zone to the *Oepikodus evae* Zone of the middle Arenig.

The *Prioniodus* sp. that is found in the Catchers Pond Group shows some differences from the typical *P. elegans*, which is found in the Cow Head Group of western Newfoundland, although it is a closely related species. The same species, *P. elegans*, found in the Catchers Pond Group, is also found in the Middle Arm Point Formation of the Humber Arm Supergroup (K. Roy, personal communication, 1989). A similar evolutionary trend is also seen in the Prioniodontacean conodonts of the Emmanuel Formation of western Australia (McTavish, 1973).

Cutwell Group

To correlate the thrust sheets of the Cutwell Group, fifty samples were collected and processed for conodonts from bedded limestones, calcareous concretions in black shales, limestone blocks in tuff breccias and blocks in limestone debris-flows, from localities shown on Figures 1 and 2.

The oldest conodonts have been extracted from blocks of limestone breccia from Limestone Island–a small rock offshore Little Bay Island (locality F10, Figure 1). This fauna, which includes a primitive *Periodon aculeatus* (Hadding) associated with *Paroistodus* cf. *P. originalis* (Sergeeva), is dominated by simple cone species and is probably late Arenig in age.

Samples from the bedded limestone at the base of the quarry on the Lushes Bight to Beaumont road (locality F4, Figure 2), blocks from limestone debris-flows from the same quarry and limestone from localities F7 and F8 (Figure 2), contain a fauna from the lower part of the *Eoplacognathus? variabilis–Microzarkodina ozarkodella* subzone of the *E.? variabilis* Zone (Table 1).

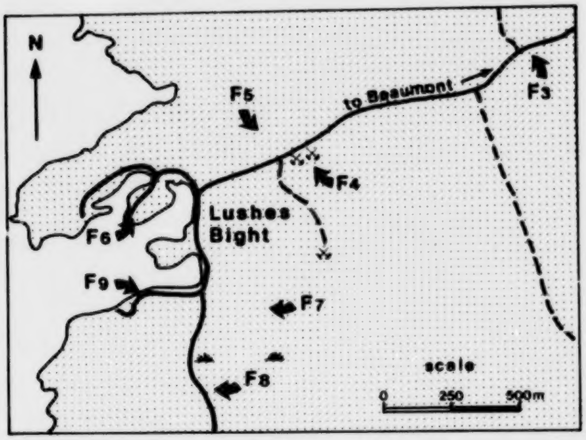

Figure 2. *Conodont sample localities in the vicinity of Lushes Bight, Long Island.*

The rest of the samples, except those from localities F1 (Figure 1) and F5 and F6 (Figure 2), come from the lower part of the *Histiodella kristina* Zone (Stouge, 1984), which is the lower part of the *Eoplacognathus suecicus* Zone. The fauna typically includes *Histiodella holodentata* Ethington and Clark, and *H. kristina* Stouge, *Cordylodus? horridus* Barnes and Poplawski, *Ansella jemtlandica* (Löfgren) and various simple cones.

Samples from localities F5 and F6 (Figure 2), and possibly F1 (Figure 1), contain the youngest fauna from the upper part of the *E. suecicus* Zone. This assemblage includes *Protoprioniodus* cf. *P. simplissimus* McTavish, *Eoplacognathus suecicus* Bergström and *Protopanderodus* cf. *P. giganteus* (Sweet and Bergström).

In some of these samples, namely those from localities F4, F6, F7 and F9 (Figure 2), a rich microfauna of inarticulate brachiopods is found in association with the conodonts. An extensive macrofauna has also been collected from different localities within the Cutwell Group. On Little Bay Island (locality F10, Figure 1) blocks of limestone contain not only conodonts but also a rich fauna of sponges, cephalopods and (?)algae. Blocks from the quarry, locality F4, and locality F5 (Figure 2) contain cephalopds, sponges and crinoid ossicles. The outcrop at locality F7 (Figure 2), is formed of a crinoidal bioclastic limestone.

Some of the conodont faunas of the Cutwell Group were obtained from the same sequences where graptolites have been described (Williams, *this volume*). These have been shown to be from the *I. victoriae maximus* and the *G. dentatus* zones of the upper Arenig and lower Lanvirn. The youngest samples contain conodonts, which in Scandinavia, are found in the Middle Ordovician *D. murchisoni* graptolite Zone.

The conodont fauna from the Cutwell Group are similar to faunas described from North America by Stouge (1984),

Table 1. Correlation of European and Table Head conodont zones with British and Newfoundland graptolite zones; ages of samples from the Catchers Pond and Cutwell groups are shown.

SYSTEM		Series (British)	British / Newfoundland Graptolite Zones	European Conodont Zones	European Conodont Subzones	Table Head Conodont Zones		Catchers Pond	Cutwell
ORDOVICIAN	MIDDLE	Lland	*G. teriusculus*	*P.serra*	*E.foliaceus*				
		Llanvirn	*D. murchisoni*	*E.suecicus*	*E.suecicus–P.sulcatus*	*H.kristina*	*Cornuodus longibasis–*	- -	- - ✕
					E.suecicus–S.gracilis		*Walliserodus ethingtoni*		✕
			G. dentatus	*E.? variabilis*	*E.?variabilis–M.ozarkodella*	*Histiodella holodentata*	*Parapanderodus arcuatus–?Erraticodon belticus*		✕
							Trigonodus carinatus–Eoneoprioniodus n.sp 1		
			U.austrodentatus		*E.?variabilis–M.flabellum parva*				
	LOWER	Arenig	*I.v.maximus*	*M.flabellum parva*		Not zoned			✕
			I.v.victoriae	*P. originalis*					
			I.v.lunatus	*B. navis*					
			D.bifidus	*O. evae*				✕	
			P.friuticosus	*P.elegans*				✕	

Barnes and Poplawski (1973), Harris *et al.* (1979), and from Scandinavia by Löfgren (1978) and Lindström (1971, 1984). The Cutwell Group, however, contains conodonts that have not been previously described or that vary only slightly from the original description of the species. It is possible that these taxa are either older or younger than those described from environments on the craton of North America or in Scandinavia. This is a characteristic of faunas from island-arcs (Dzik, 1982; Lindström, 1984; Neuman, 1984).

The age of the conodont fauna from a shale sequence on Oil Island also has been constrained by a 469 $^{+5}_{-3}$ Ma U–Pb zircon age of felsic tuff overlying conodont-bearing shales (Dunning and Krogh, 1988), which is approximately the Arenig/Llanvirn boundary on the van Eysinga (1975) time table.

CONCLUSIONS

The difference in age of the samples from the Catchers Pond Group has shown that the limestone bed is younger toward the north and time of its deposition spans the early to middle Arenig.

The oldest conodonts from the Cutwell Group probably represent the boundary of the *Microzarkodina flabellum parva* Zone and the *Eoplacognathus? variabilis* Zone. The main body of the fauna is from the top of the *E.? variabilis* Zone to the lower part of the *Eoplacognathus suecicus* Zone. The youngest fauna comes from the *E. suecicus–P. sulcatus* subzone, the upper part of the *E. suecicus* Zone. This spans the late Arenig to at least the late Llanvirn, and possibly to the base of the Llandeilo.

The conodonts from the Cutwell Group show that there is a slight difference in age of the different thrust slices, which also corresponds with the results from the graptolites (Williams, *this volume*). It is not clear at the moment whether any of the rocks of the Cutwell Group are as young as the base of the Llandeilo. It is possible that these taxa appear earlier in this island-arc environment than elsewhere. Both early and late appearances have been described for brachiopod genera from Iapetus by Neuman (1984). Some taxa also appear earlier in the Upper Ordovician Tasman island-arc (B.D. Webby, personal communication, 1988).

The 469 $^{+5}_{-3}$ Ma U–Pb zircon age obtained for the felsic volcanic rocks of the Cutwell Group is coeval with 473 $^{+3}_{-2}$ Ma U–Pb zircon age reported for felsic volcanics of the Buchans Group (Dunning and Krogh, 1988). The conodont assemblage extracted from limestone blocks of the Buchans Group (Nowlan and Thurlow, 1984, 1987) is also similar to that of the main part of the Cutwell Group. As well, conodonts from both groups display a similar CAI of 4 to 4.5, which suggests that they have been heated to 300° C (Epstein *et al.*, 1977; Rejebian *et al.*, 1987). The similarities in U–Pb zircon age, conodont assemblage and CAI strongly suggest that the Cutwell and Buchans groups are close correlatives (O'Brien and Szybinski, 1988).

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A PROPOSAL FOR REVISED STRATIGRAPHIC NOMENCLATURE OF THE GANDER AND DAVIDSVILLE GROUPS AND THE GANDER RIVER ULTRABASIC BELT, OF NORTHEASTERN NEWFOUNDLAND

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Newfoundland Mapping Section

ABSTRACT

The Weir's Pond map area, centred 30 km northeast of Gander, contains parts of two major tectonostratigraphic divisions in northeastern Newfoundland. The Dunnage Zone, to the northwest, is represented by the Ordovician Davidsville Group and the Ordovician or older Gander River complex; the Gander Zone, to the southeast, is represented by the Ordovician or older Gander Group. The Davidsville Group is subdivided into three formations and the Gander Group is divided into two formations. The Gander River complex, is structurally imbricated with the basal part of the Davidsville Group.

REGIONAL RELATIONSHIPS

The northeastern corner of the Weir's Pond map area straddles the contact between the two major tectonostratigraphic divisions of the Appalachian Central Mobile Belt, i.e., the Gander Zone and the Dunnage Zone (Figure 1).

The Early Ordovician and/or earlier Gander Group (Kennedy and McGonigal, 1972), representing the Gander Zone (Williams, 1976), underlies much of the central and eastern parts of the area, and has been mapped northeastward to the Carmanville (2E/8) map area (Currie and Pajari, 1977) and southwestward, for 100 km, to the Baie d'Espoir area (Dickson, 1987). The Gander Group passes gradationally eastward into amphibolite-facies metasedimentary rocks, and is interpreted to be the protolith of the Hare Bay Gneiss (Blackwood, 1978).

A package of undated ultramafic, mafic and felsic rocks, the Gander River complex (the Gander River Ultrabasic Belt of Jenness, 1958; Blackwood, 1982), in places, bounds the Gander Group to the west and defines the eastern margin of the Dunnage Zone. The complex extends discontinuously northeastward to the coast, and occurs sporadically south of Gander Lake.

The Davidsville Group (Kennedy and McGonigal, 1972) forms the basal sedimentary division in the eastern part of the Dunnage Zone. Southward, the group passes along strike into the Baie d'Espoir Group (Colman-Sadd, 1980; Dickson, 1987) and northeastward, the group has been mapped to the coast (Currie *et al.*, 1980). The basal part of the Davidsville Group is structurally imbricated with the Gander River complex.

LOCAL GEOLOGY

Gander Group

The Gander Group has been informally divided into the Jonathans Pond and Indian Bay formations (O'Neill and Knight, 1988). The term Indian Bay formation was first introduced by Wonderley (Wonderley and Neuman, 1984) but redefined by O'Neill (O'Neill and Knight, 1988). It is proposed (O'Neill, *in preparation)* to amend the term Indian Bay formation to Indian Bay Big Pond formation. This revision is necessary in order to accurately locate the formation, which is exposed principally on the shores, and in the vicinity, of Indian Bay Big Pond. Indian Bay itself, is located on the coast about 15 km to the east. Brief lithological descriptions of the two formations are given below (a detailed account is presented in O'Neill and Knight, 1988 and O'Neill, *in preparation).*

The Jonathans Pond formation. The Jonathans Pond formation consists primarily of thin to thick interbedded psammite, semipelite and pelite; it also includes quartz-granule sandstone, quartz arenite, calc-silicate layers or lenses, and mafic sills or dykes.

The Indian Bay Big Pond formation. The Indian Bay Big Pond formation is a northeast-striking sequence of metasedimentary and metavolcanic rocks exposed in a narrow belt in the southeastern corner of the Weir's Pond map area. The sedimentary rocks include locally interbedded, grey to dark-grey, buff and maroon sandstone, green, maroon and black pelite and a 6-m-thick chert band. Fossiliferous, fine-grained sandstone boulders and conglomerate boulders, typically angular, are abundant locally and appear to be restricted to the shores of Indian Bay Big Pond. The fossil-

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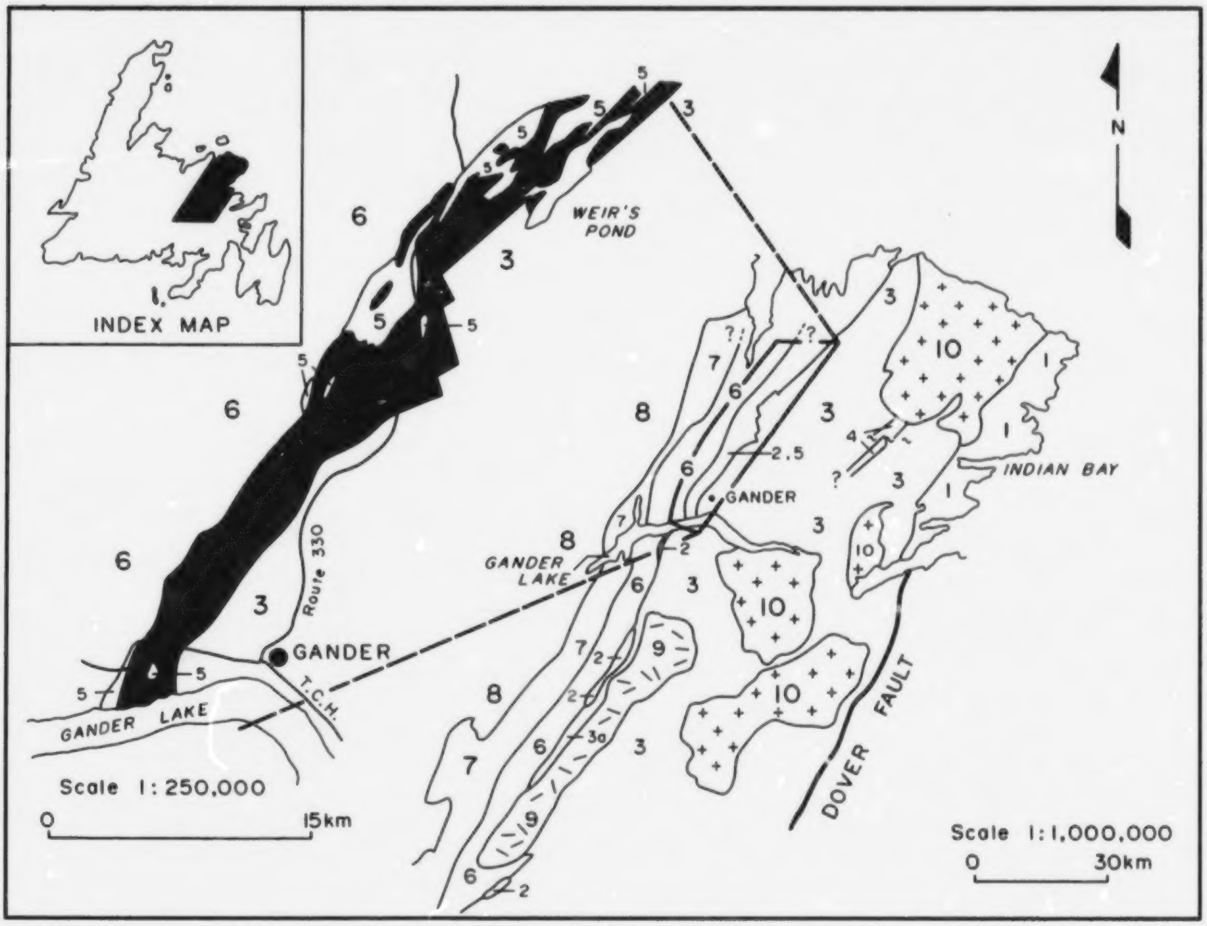
This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989


INDEX MAP
N
WEIR'S POND
Route 330
GANDER
T.C.H.
GANDER LAKE
Scale 1:250,000
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1
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LEGEND

10 DEVONIAN GRANITE
9 SILURO – DEVONIAN GRANITE

SILURIAN

8 BOTWOOD GROUP

ORDOVICIAN

DAVIDSVILLE GROUP (Units 5 to 7)

7 THE OUTFLOW FORMATION
6 HUNTS COVE FORMATION
5 WEIRS POND FORMATION

EARLY ORDOVICIAN OR EARLIER

GANDER GROUP (Units 3 and 4)

4 INDIAN BAY BIG POND FORMATION
3 JONATHANS POND FORMATION
2 **GANDER RIVER COMPLEX**

SILURO – DEVONIAN AND OLDER

1 UNSEPARATED HARE BAY GNEISS AND GRANITOID ROCKS

Figure 1. *Location map and geology of northeastern Newfoundland showing the revised stratigraphic nomenclature.*

rich boulders are assumed to be derived from a lithologically similar unit within the Indian Bay Big Pond formation. The fauna indicate a late Arenig age (Wonderley and Neuman, 1984). Not all contacts between units in this formation are exposed, and therefore the complete internal stratigraphic continuity of the formation is only assumed. Stratigraphic indicators are lacking in the formation and the relative ages of the units are unknown.

Sandstone of the Indian Bay Big Pond formation is typically much richer in lithic detritus than those of the Jonathans Pond formation. One common type of lithic clast in the Indian Bay Big Pond formation is fine- to medium-grained muscovite ± biotite schist. These fragments usually feature one foliation that is clearly crenulated in some clasts, suggesting that the source terrain included polydeformed, metamorphosed (at least to greenschist facies) rocks. An isotropic mineral has been tentatively identified as detrital garnet in some thin sections. Volcanic clasts also occur in the sandstones.

The Indian Bay Big Pond formation is in gradational contact to the northwest with the Jonathans Pond formation. At the contact, grey sandstone and grey-green pelite, of the Jonathans Pond formation, are interbedded with dark-grey sandstone and black laminated pelite of the Indian Bay Big Pond formation. The proportion of black pelite increases eastward and forms about 50 percent of the exposure, near the eastern boundary of this gradational unit. Southward, the contact between the two formations, i.e., on the northwest side of Indian Bay Big Pond, is poorly exposed. However, the rocks in both formations are lithologically similar, and the contact is interpreted to be conformable.

The southeast contact of the Indian Bay Big Pond formation is more problematic because of poor exposure and the presence of a several-km-wide structural high-strain zone around Wing Pond. North of Indian Bay Big Pond, rocks, including some within the Indian Bay Big Pond formation, are heterogeneously highly strained. The contact between the two formations has been modified by this high-strain zone and therefore the nature of the original boundary is unknown. The southeastern belt of the Jonathans Pond formation is interpreted to be a structural repetition.

GANDER RIVER COMPLEX

O'Neill *(in preparation)*, formally proposes to redefine the term Gander River Ultrabasic Belt and amend it to the Gander River complex. Much of the complex comprises gabbroic and mafic volcanic–volcaniclastic rocks; felsic plutonic rocks form a minor but significant component. Also, most, if not all, of the unaltered ultramafic rock in the complex is pyroxenite or pyroxene rich and is therefore relatively SiO_2 rich.

The Gander River complex has not been dated. However, overlying fossiliferous sedimentary rocks containing abundant ultramafic, mafic and felsic detritus (presumeably eroded from the complex) range from late Arenig to lower Llandeilo and provide a minimum age for the complex.

DAVIDSVILLE GROUP

O'Neill *(in preparation)*, proposes to subdivide the Davidsville Group into three formations; the subdivision and new nomenclature are briefly discussed here. The eastern formation is termed the Weir's Pond formation and is readily extrapolated southwestward to the north shore of Gander Lake. The Davidsville Group west of the Weir's Pond formation, has been divided into two units (Blackwood, 1979, 1980). These two units define the central and western formations and are termed the Hunt's Cove and The Outflow formations respectively.

The Weir's Pond formation. The name Weir's Pond formation is proposed for fossiliferous sandstone, conglomerate, fossiliferous limestone and graphitic black shale units that are structurally imbricated with the Gander River complex and are well exposed in the Weir's Pond area. The oldest rock type in the Weir's Pond formation is a fine- to medium-grained, grey, fossiliferous arenite containing late Arenig fauna of mixed North American and Baltoscandian (European) aspect (Boyce *et al.*, 1988). The rock also contains abundant chromite detritus. This unit is not seen in contact with other units of the formation but interbedded conglomerate, sandstone and siltstone is inferred to lie stratigraphically above the fossiliferous sandstone. This relationship is based on correlation with rocks on the north shore of Gander Lake, several kilometres west of Little Harbour, where fossiliferous rocks of late Arenig age pass conformably westward into conglomeratic sandstones, locally.

On the northwest side of Weir's Pond, fossiliferous limestone nonconformably overlies serpentinized ultramafic rock (Blackwood, 1982) of the Gander River complex. The fauna collected from the limestone are of mixed North American and Baltoscandian aspect, Llanvirn to Llandeilo in age, and similar to those recognized in the Cobbs Arm Limestone, Summerford Group on New World Island and in the Spruce Brook Formation of the Through Hill area (Boyce *et al.*, 1988).

Caradoc graphitic black shale of the Weir's Pond formation is exposed near the Middle Ordovician limestone, but both are separated by an exposure gap.

Hunt's Cove formation. The name Hunt's Cove formation is informally proposed for a thick sequence of slate and minor conglomerate, mostly outcropping in the Gander River area. The name is taken from a cove on the south shore of Gander Lake where sections of this formation are well exposed. The Hunt's Cove formation is exposed in the northwestern corner of the Weir's Pond area, near Gander River and it is inferred to be in stratigraphic continuity with the Weir's Pond formation. The formation continues southwestward, through map areas 2E/2, 2D/15, 2D/10, 2D/11, 2D/6 to 2D/3. In map area 2D/3, similar rocks are referred to as the St. Joseph's Cove Formation of the Baie d'Espoir Group. Detailed work is required in the area of the transition to substantiate this equivalence. The St. Joseph's Cove Formation was defined by Colman-Sadd (1976) for pelite, siltstone, sandstone and volcanic rocks in the St Alban's map area; it was mapped in the 2D/3 map area by Dickson (1987).

The Outflow Formation. The name The Outflow formation is informally proposed for the westernmost unit of the Davidsville Group. Some of the best exposures of the formation occur on the north shore of Gander Lake where 'the Outflow' into the Gander River occurs. It comprises fine- to coarse-grained sandstone, locally with shale intraclasts, interbedded with grey to black siltstone and slate; conglomeratic beds form a minor component. West of 'the Outflow' on Gander Lake, a conformable but abrupt contact occurs with the Hunt's Cove formation. The Outflow formation is mappable northeastward to the coast and southwestward through map areas 2E/2, 2D/15, 2D/14, 2D/11, 2D/6 to 2D/5. In map area 2D/5, similar rocks are referred to as the North Steady Pond Formation (Colman-Sadd, 1980). Again detailed mapping is required to substantiate this equivalence. The North Steady Pond Formation comprises greenish grey pelite, siltstone and sandstone, as well as minor conglomerate and mafic to felsic volcanic rocks; it grades laterally into the St. Josephs Cove Formation (Colman-Sadd, 1976). Dickson (1987), however, in the Mount Sylvester area, interprets the contact between the St. Joseph's Cove and North Steady Pond formations to be a fault, locally marked by ultramafic lenses.

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TECTONOTHERMAL HISTORY AND ^{40}AR/^{39}AR GEOCHRONOLOGY OF NORTHEASTERN GANDER ZONE, WEIR'S POND AREA (2E/1)

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ABSTRACT

Two phases of deformation, manifested by mesoscopic, isoclinal and later upright, open folds, were identified in the map area. The affects of three types of metamorphism, apparently developed after the initial phase of deformation, are described.

Biotite, garnet, staurolite, andalusite, sillimanite and locally cordierite isograds, are defined for the northeastern part of the map area, near Ocean Pond. Microfabric–porphyroblast relationships imply overlap of metamorphic mineral growth and open upright folding. The sequence of metamorphic mineral assemblages developed, suggest low pressure–high temperature regional metamorphism.

In the southeastern part of the map area, near Wing Pond, a series of isograds, which are at least locally telescoped, occur within a structural high-strain zone. Kyanite-bearing assemblages are present and imply higher pressure metamorphism (deeper level?) than that at Ocean Pond.

Coarse-grained muscovite, overprinting all earlier structures and minerals, occurs in a narrow zone, in the northeast, peripheral to the Deadman's Bay Granite. Near Moccasin Pond, cordierite–andalusite–K-feldspar assemblages are developed within 2 km of the Deadman's Bay Granite. These assemblages define a zone of hornfelsing associated with this granite.

^{40}Ar/^{39}Ar ages on muscovite and biotite concentrates from pelitic rocks near Ocean Pond, range between 400 and 385 Ma, and define a cooling age for the regional metamorphism in the northeast. The time frame for this thermal event suggests a correlation between metamorphism and plutonism in northeastern Newfoundland, and a Late Silurian orogeny in southern Newfoundland.

INTRODUCTION

Regional tectonic and stratigraphic relationships of the rocks discussed in this paper are described elsewhere (O'Neill and Blackwood, *this volume*). This paper presents an account of microfabric–porphyroblastesis relationships in metasedimentary rocks of the Gander Group and ^{40}Ar/^{39}Ar cooling ages from muscovite and biotite mineral separates, which can be used to infer the approximate timing of the development of some of these relationships.

TECTONOTHERMAL HISTORY

General Description

Gander Group rocks in the Weir's Pond map area exhibit the effects of three types of metamorphism. To facilitate the description of these metamorphic events, the part of the map area underlain by rocks at biotite and higher grades is divided into three areas, the Ocean Pond, Moccasin Pond and Wing Pond areas. However, this report primarily describes the tectonothermal history of the rocks in the Ocean Pond area, since most of the ^{40}Ar/^{39}Ar data presented here, relate to the area (Figure 1). A hornfelsed aureole, peripheral to the Deadman's Bay Granite, is also described.

Two phases of deformation were recognized on a mesoscopic scale in the map area. The earliest phase of deformation produced isoclinal folds having a predominantly steep, westerly to southwesterly plunge and a north–northeast-trending axial planar fabric. These early-phase folds are overprinted by a later phase of open, upright folds having a northeast-trending subhorizontal plunge and a typically steep crenulation cleavage.

Ocean Pond Area

In the Ocean Pond area, biotite, garnet, staurolite, andalusite and sillimanite isograds are defined in rocks of the Jonathans Pond formation. Pelitic and semipelitic rocks are coarse-grained and schistose throughout much of the zone. The sillimanite isograd defines a localized, metamorphic culmination near Ocean Pond. Farther east and northeast,

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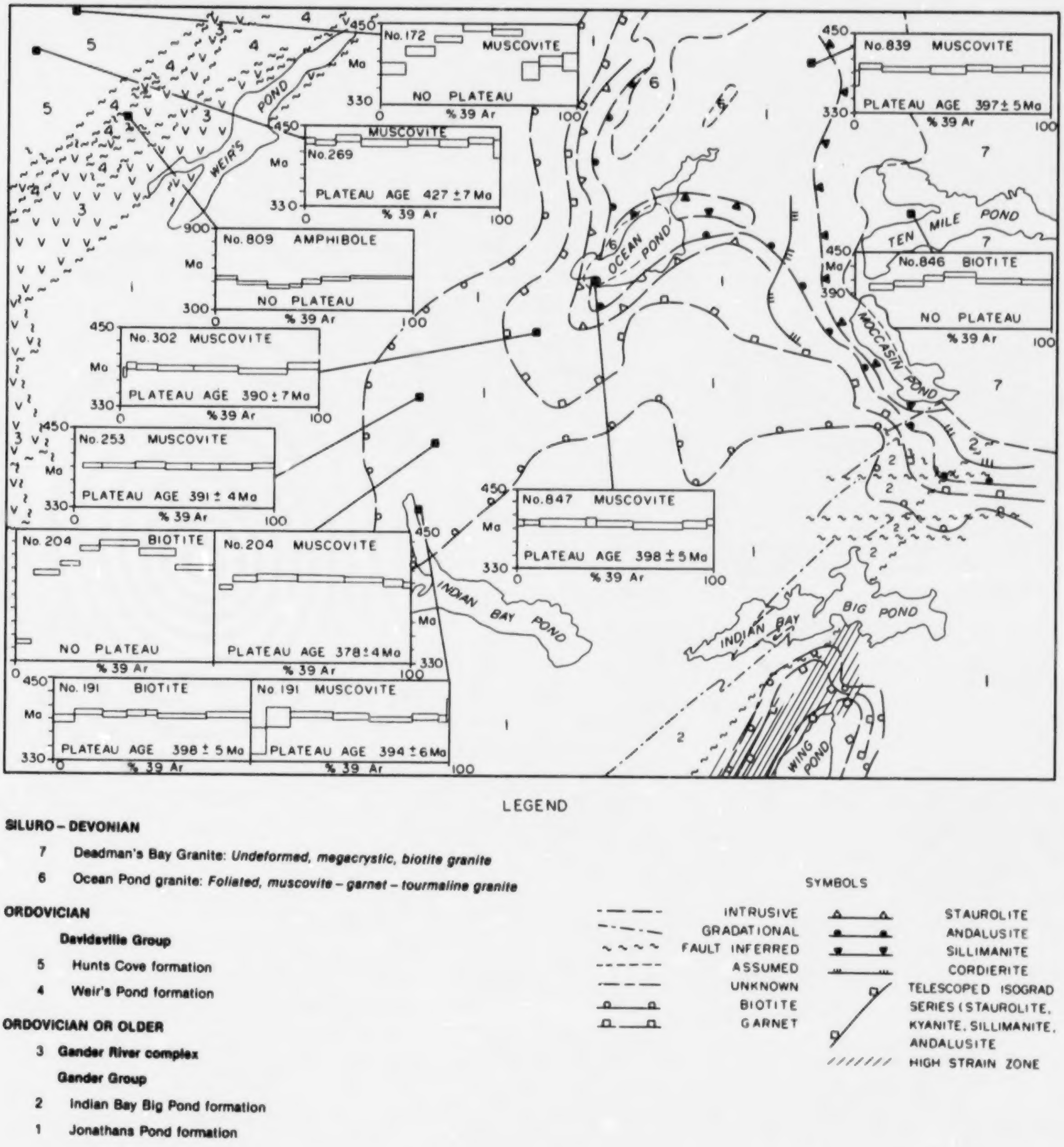
No. 172 MUSCOVITE
NO PLATEAU
No. 269 MUSCOVITE
PLATEAU AGE 427 ± 7 Ma
No. 809 AMPHIBOLE
NO PLATEAU
No. 302 MUSCOVITE
PLATEAU AGE 390 ± 7 Ma
No. 253 MUSCOVITE
PLATEAU AGE 391 ± 4 Ma
No. 204 BIOTITE
NO PLATEAU
No. 204 MUSCOVITE
PLATEAU AGE 378 ± 4 Ma
No. 191 BIOTITE
PLATEAU AGE 398 ± 5 Ma
No. 191 MUSCOVITE
PLATEAU AGE 394 ± 6 Ma
No. 847 MUSCOVITE
PLATEAU AGE 398 ± 5 Ma
No. 839 MUSCOVITE
PLATEAU AGE 397 ± 5 Ma
No. 846 BIOTITE
NO PLATEAU
Ma
% 39 Ar
WEIR'S POND
OCEAN POND
TEN MILE POND
MOCCASIN POND
INDIAN BAY POND
INDIAN BAY BIG POND
WING POND
LEGEND
SILURO – DEVONIAN
7 Deadman's Bay Granite: Undeformed, megacrystic, biotite granite
6 Ocean Pond granite: Foliated, muscovite – garnet – tourmaline granite
ORDOVICIAN
Davidsville Group
5 Hunts Cove formation
4 Weir's Pond formation
ORDOVICIAN OR OLDER
3 Gander River complex
Gander Group
2 Indian Bay Big Pond formation
1 Jonathans Pond formation
SYMBOLS
INTRUSIVE
GRADATIONAL
FAULT INFERRED
ASSUMED
UNKNOWN
BIOTITE
GARNET
STAUROLITE
ANDALUSITE
SILLIMANITE
CORDIERITE
TELESCOPED ISOGRAD SERIES (STAUROLITE, KYANITE, SILLIMANITE, ANDALUSITE
HIGH STRAIN ZONE

Figure 1. *Geology and $^{40}Ar/^{39}Ar$ mineral age spectra for the Weir's Pond map area, northeastern Gander Zone, Newfoundland.*

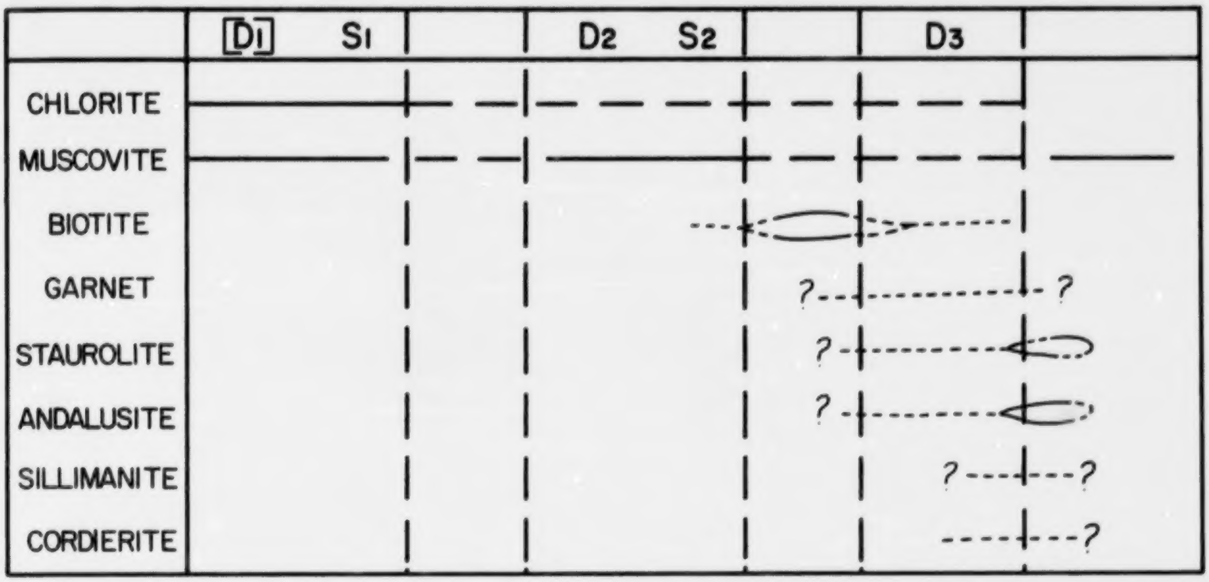

Figure 2. *Relative timing of metamorphic mineral growth and relationship to deformational phases; Ocean Pond area, Weir's Pond map area.*

the sillimanite isograd approximately parallels the margin of the Deadmans Bay Granite, from the northern edge of the map area, southward to Moccasin Pond, where the granite intrudes across the isograd.

The leucocratic, muscovite-, garnet- and tourmaline-bearing Ocean Pond granite is best exposed near Ocean and Little Ocean ponds, in the higher grade parts of the Ocean Pond area. Although the granite appears to cut across some of the isograds, it does not have an associated aureole.

Definition of fabrics. In most of the metasedimentary rocks, the principal tectonic fabric is a penetrative foliation and is axial planar to isoclinal folds. This fabric is manifested as a solution seam cleavage in psammite and as a slatey cleavage in pelite. However, microfabric studies, particularly on pelitic and semipelitic rocks, show that this principal fabric is almost invariably a crenulation of an older fabric, which is the oldest recognized and therefore designated S_1. No major deformational phase has been associated with S_1 and it may be a bedding-parallel layering. The principal fabric, S_2, is therefore S_1 incompletely transposed. S_2 is axial planar to the isoclinal folds that occur in most exposures. In most semipelitic and pelitic samples, S_2 is gently to moderately crenulated by a later folding, designated F_3. These late microscopic crenulations correlate with the ubiquitous, open folds and crenulations on a mesoscopic scale. These folds are northeast-trending and typically display a subhorizontal plunge.

Fabric development and porphyroblastesis. Both S_1 and S_2 fabrics are defined by fine- to medium-grained chlorite and muscovite, which gradually increase in grain size with increasing grade. Biotite poikiloblasts overprint S_1 and S_2 and contain inclusions of these fabrics. The relationship between biotite growth and S_3 is more subtle. Biotite never clearly overprints the F_3 folds, however, toward the margin, of approximately 20 percent of the biotite examined, Si (internal fabric) in biotite, (which is typically S_2) is bent and parallels a corresponding curvature of the S_2 fabric in the matrix, caused by F_3 folds. Elsewhere, biotite is bent around F_3 folds. Thus, biotite is post-D_2 but grew before, and synchronously with, the early stages of D_3 (Figure 2).

Almanditic garnets overprint biotite and typically contain a poorly developed Si (typically S_2), which is generally straight but locally gently curved.

Most staurolite and andalusite porphyroblasts contain an Si defined by quartz, opaque minerals and muscovite. They also contain random biotite, garnet and occasionally coarse grained muscovite. Inclusion trails in staurolite and andalusite are straight, curved or crenulated. As in the case of biotite, staurolite and andalusite clearly overprint the S_1 and S_2 fabrics. However, unlike biotite, staurolite and andalusite overprint F_3 crenulations locally. The crenulations preserved in the staurolite and andalusite are continuous with F_3 crenulations defined in the matrix. There is no apparent flattening of the fabrics around the porphyroblasts and, therefore, andalusite and staurolite grew after D_3. Although porphyroblast growth may have been partly synchronous with the development of D_3, there is no conclusive evidence for this.

Fibrolite−sillimanite developed after and replaces andalusite and, therefore, probably has similar timing relationships to D_3 as andalusite. However, a high-strain fabric involving sillimanite is developed in several samples collected northwest of Ten Mile Pond. This high-strain fabric is not readily correlated with any of the previously defined fabrics. Both fibrolite and sillimanite are aligned in foliae that are evident in hand specimen and are clearly crenulated. The sillimanite foliae are variable in width, up to a maximum of 1 mm; the thicker foliae split and anastomose around single or aggregate quartz grains and biotite. Some of the quartz grains enveloped by sillimanite have aspect ratios of 10:1 to 15:1, and are en echelon and strained. Biotite enveloped by the aluminosilicate is locally fish-shaped, suggesting a non-coaxial deformation. In contrast to the strong morphology exhibited by the sillimanite within the foliae, sillimanite within quartz grains away from the foliae are randomly oriented. Textures involving sillimanite, which are similar to those described above, have been described by Vernon (1987), who attributes their development to the ability of sillimanite to undergo non-coaxial deformation.

Moccasin Pond Area

The textural characteristics of the metasediments, from the northern end of Moccasin Pond southward, contrast with those exhibited by the rocks in the Ocean Pond area. In the Moccasin Pond area, biotite, garnet, andalusite and cordierite isograds are defined in pelitic and semipelitic rocks of the Jonathans Pond and Indian Bay Big Pond formations. The isograds approximately parallel the margins of the Deadman's Bay Granite. The principal differences between rocks of the Ocean Pond and Moccasin Pond areas are in grain size and in their microfabric–porphyroblastesis relationships. Texturally, the rocks in the Moccasin Pond area, which are biotite and garnet grades, are fine grained. At higher grades, andalusite and cordierite porphyroblasts, up to 4 mm across, occur in a fine grained matrix. Andalusite and cordierite porphyroblasts have, locally, survived the extensive retrogression affecting these rocks, although most are partly sericitized.

Fabric development and porphyroblastesis. The earliest fabric in pelitic and semipelitic rocks of the Moccasin Pond area, is poorly developed and defined by very fine grained, chlorite and muscovite. In those rocks where this early tectonic? fabric is preserved, fine biotite randomly overprints it, and the biotite is in turn overprinted by pseudomorphs of fine grained white mica and chlorite after cordierite and/or andalusite(?). After the porphyroblast growth, the rocks are affected by a second deformation, which crenulated the early fabric and aligned the biotite. The new crenulation fabric is particularly strong around pseudomorphs and porphyroblasts of andalusite and cordierite, which form augen. However, locally, the Si in the pseudomorphs curve at their margins and become parallel with the external matrix fabric. Elsewhere the Si is moderately crenulated, thus implying growth of these porphyroblasts early in the crenulation deformation. This deformation later affected the augening of the porphyroblasts. The biotite fabric is locally overprinted by garnet and a late crenulation gently folds the biotite fabric in some sections.

Wing Pond Area

Near Wing Pond, a rapid increase in metamorphic grade occurs several kilometres southeast of the southeastern margin of the Indian Bay Big Pond formation. In the lower grade parts of the zone, the rocks are coarse grained and muscovite-rich. The biotite zone is based on rare and sporadic occurrences of this mineral. Similarly, the garnet ± biotite zone, is based on several scattered occurrences of this assemblage. On the northeast side of the pond, the metastable assemblage kyanite–staurolite–sillimanite–andalusite–biotite ± garnet occur within several hundred metres of rocks containing biotite and garnet. The kyanite and staurolite commonly occur as inclusions in andalusite porphyroblasts and form a relict, metastable assemblage. It is not clear whether the sillimanite is associated with the older, kyanite–staurolite assemblage or the andalusite-bearing assemblage.

The close spatial proximity of the lower and higher grade rocks suggest that several isograds, including staurolite, andalusite and sillimanite are telescoped together. The surface traces of all the isograds in the Wing Pond area, lie within a steep high-strain zone. Rocks in this zone, containing the assemblages described above, all exhibit the effects of non coaxial deformation.

Kyanite-bearing rocks indicate a higher pressure type of metamorphism than that in the Ocean Pond area; however, this may represent a deeper level but equivalent metamorphism. The timing of the development of the high-strain zone and its relationship to the regional metamorphism of the Ocean Pond area is not known. The tectonothermal history of the Wing Pond area will be described in detail (O'Neill, *in preparation).*

Deadman's Bay Granite Aureole

In the Ocean Pond area, sillimanite-bearing rocks, in the northeast, within 3 km of the margin of the Deadman's Bay Granite, are characterized by coarse porphyroblastic muscovite, up to 3 mm across. Elsewhere, in the Ocean Pond area, away from the granite, muscovite is typically fine to medium grained ($\leq$ 0.5 mm across) and is the main fabric-forming element in S_1 and S_2 fabrics. The coarse grained muscovite in rocks adjacent to the Deadman's Bay Granite, contain inclusions of sillimanite and biotite and, therefore, grew late with respect to the regional metamorphism. Tectonic fabrics are only vaguely defined or, are totally absent from these hornfelsed rocks.

West of Ten Mile Pond, four samples, two within sillimanite and two in andalusite grade rocks, contain cordierite. In all the samples, andalusite and biotite are associated with the cordierite, but are enclosed within cordierite in some samples. In one sample, a fibrolite rim around the andalusite and the biotite is included within the

Table 1: $^{40}Ar/^{39}Ar$ analytical data on muscovites and biotites from the Weirs Pond area, northeastern Gander Zone.

Release Temp C	$(^{40}Ar/^{39}Ar)^*$	$^{37}Ar/^{39}Ar$	$(^{36}Ar/^{39}Ar)^*$	Moles ^{39}Ar	^{39}Ar % of Total	%^{40}Ar Radiogenic	K/Ca	Apparent Age(Ma)
				MUSCOVITE				
Sample #191; Pelite; Gander Group; J = .006113								
825	51.45	0.4392	0.0531	16.3	7.0	69.5	1.12	356.7 +/-19.6
890	43.32	0.4218	0.0117	29.3	12.6	92.0	1.16	393.5 +/-13.7
950	41.22	0.2877	0.0033	49.4	21.3	97.6	1.70	396.7 +/-4.0
1025	40.53	0.2581	0.0018	43.6	18.8	98.6	1.90	394.5 +/-5.2
1115	40.23	0.2332	0.0020	50.1	21.6	98.5	2.10	391.3 +/-4.3
1170	40.15	0.4186	0.0004	29.3	12.6	99.7	1.17	395.0 +/-4.0
1220	40.43	0.8757	0.0024	10.4	4.5	98.3	0.56	392.6 +/-4.9
Fusion	42.46	1.1932	0.0039	3.3	1.4	97.4	0.41	406.9 +/-16.3
Total				231.4	100.0		Total Gas Age	391.6 +/-6.8
Sample #204; Pelite; Gander Group; J = .005926								
825	34.62	0.0147	0.0120	40.3	2.8	89.7	33.27	304.6 +/-8.1
890	39.10	0.0300	0.0040	91.6	6.5	96.9	16.35	365.3 +/-3.6
950	39.95	0.0204	0.0020	168.4	11.9	98.4	24.07	377.9 +/-4.0
1025	39.98	0.0101	0.0015	296.4	21.0	98.8	48.47	379.5 +/-3.8
1115	39.85	0.0144	0.0015	334.8	23.7	98.8	33.93	378.3 +/-3.8
1170	39.68	0.0204	0.0013	270.4	19.1	98.9	24.03	377.4 +/-3.9
1220	39.22	0.0495	0.0016	148.2	10.5	98.7	9.90	372.6 +/-3.7
Fusion	38.88	0.0630	0.0016	64.4	4.5	98.7	7.77	369.7 +/-4.5
Total				1414.4	100.0		Total Gas Age	374.4 +/-4.0
Sample #253; Pelite; Gander Group; J = .005992								
725	38.690	0.0065	0.0107	44.2	2.5	91.7	75.84	347.86 +/-3.69
825	41.400	0.0019	0.0030	175.5	9.9	97.8	261.33	391.72 +/-4.01
890	40.970	0.0014	0.0015	295.8	16.7	98.8	354.56	391.80 +/-3.96
950	40.850	0.0011	0.0010	284.7	16.0	99.2	428.32	392.23 +/-5.03
1025	40.760	0.0037	0.0013	235.3	13.3	99.0	132.08	390.53 +/-3.98
1115	40.660	0.0039	0.0013	249.6	14.1	99.0	126.61	389.74 +/-3.95
1170	40.670	0.0040	0.0012	295.8	16.7	99.1	110.63	390.16 +/-3.93
Fusion	41.200	0.0015	0.0019	195.0	11.0	98.6	336.31	392.99 +/-3.98
Total				1775.8	100.0		Total Gas Age	390.17 +/-4.13

cordierite; however, timing relationships between the cordierite and sillimanite blastesis are equivocal.

K-feldspar is present in stable coexistence with the cordierite−sillimanite assemblage.

Southwest of Big Bear Cove Pond, andalusite−biotite assemblages also contain K-feldspar. The presence of K-feldspar implies that the stability field of muscovite + quartz has been exceeded and the rocks are high grade.

The spatial distribution of these assemblages in hornfelsed rocks, peripheral to this granite, indicates that the zone represents the granite aureole.

^{40}AR/^{39}AR GEOCHRONOLOGY

$^{40}Ar/^{39}Ar$ incremental release ages were determined for 8 muscovite, 3 biotite and 1 hornblende mineral concentrate, from metasedimentary and igneous rocks in the Weir's Pond area. The analytical data are presented in Table 1 and the

Table 1: *(Continued)*

Release Temp C	(^{40}Ar/^{39}Ar)*	^{37}Ar/^{39}Ar	(^{36}Ar/^{39}Ar)*	Moles ^{39}Ar	^{39}Ar % of Total	%^{40}Ar Radiogenic	K/Ca	Apparent Age(Ma)
Sample #839; Sillimanite Schist; Gander Group; J = .006172								
825	51.97	0.0115	0.0469	28.0	1.8	73.3	42.57	380.8 +/-11.3
890	42.19	0.0060	0.0069	182.0	11.6	95.1	81.26	399.2 +/- 4.3
950	40.76	0.0071	0.0034	392.6	25.1	97.5	69.47	395.6 +/- 4.1
1025	40.68	0.0021	0.0032	286.0	18.3	97.6	239.02	395.3 +/- 5.3
1120	40.98	0.0036	0.0033	201.5	12.9	97.6	136.68	397.9 +/- 4.3
1170	40.68	0.0084	0.0027	244.4	15.6	98.0	58.66	396.8 +/- 4.0
1220	40.67	0.0035	0.0024	144.3	9.2	98.2	141.70	397.4 +/- 5.2
Fusion	40.83	0.0963	0.0030	83.9	5.4	97.7	5.09	397.2 +/- 4.4
Total				1562.6	100.0		Total Gas Age	396.4 +/- 4.6
Sample #302; Pegmatite; Ocean Pond Granite; J = .006054								
750	45.04	0.0032	0.0197	28.6	0.8	87.0	152.74	384.0 +/-20.8
825	41.23	0.0085	0.0076	54.6	1.5	94.5	57.49	382.0 +/- 8.1
890	41.14	0.0008	0.0033	187.2	5.2	97.5	652.64	392.3 +/- 4.9
950	40.81	0.0050	0.0030	400.4	11.1	97.7	98.75	390.2 +/- 4.3
1025	40.38	0.0003	0.0020	655.2	18.2	98.4	1938.29	389.1 +/- 4.2
1115	40.24	0.0004	0.0017	821.6	22.8	98.7	1366.04	388.7 +/- 4.1
1170	40.04	0.0012	0.0019	895.7	24.8	98.5	419.52	386.2 +/- 4.1
Fusion	40.80	0.0007	0.0018	561.6	15.6	98.6	703.11	393.2 +/- 4.5
Total				3604.9	100.0		Total Gas Age	389.1 +/- 4.4
Samples #847; Ocean Pond Granite; J = .006168								
825	45.28	0.0026	0.0180	52.7	3.2	88.2	188.90	397.3 +/- 4.2
890	42.92	0.0007	0.0095	138.5	8.4	93.4	716.37	398.5 +/- 4.0
950	41.85	0.0018	0.0061	387.4	23.4	95.6	274.36	398.1 +/- 5.0
1025	41.53	0.0006	0.0048	83.9	5.1	96.5	866.49	398.6 +/- 5.5
1120	41.43	0.0010	0.0049	312.7	18.9	96.4	491.38	397.4 +/- 4.6
1170	40.88	0.0027	0.0040	411.5	24.8	97.0	184.07	394.8 +/- 4.4
1220	41.26	0.0023	0.0045	204.1	12.3	96.7	214.07	396.9 +/- 5.3
Fusion	41.40	0.0057	0.0039	67.0	4.0	97.1	86.16	399.6 +/- 4.7
Total				1657.5	100.0		Total Gas Age	397.1 +/- 4.7

technique is described briefly here. Age spectra and their locations are shown in Figure 1 for most of the muscovite and biotite separates; the remainder of the spectra for muscovite and hornblende will be discussed in O'Neill *(in preparation)*.

^{40}Ar/^{39}Ar incremental release spectra allow cooling ages to be determined for various K-bearing minerals. These ages date the approximate time the minerals passed through their closure temperatures i.e., the temperature Ar ceased to diffuse through the crystal lattice. However, in the case of rapidly cooled rocks, the crystallization and cooling ages converge. Two types of ages, plateau and total gas, are derived from the incremental release spectra. A plateau age is the mean value of all increments, which are concordant, based on 2 sigma analytical uncertainties, excluding any uncertainty in the J factor. A total gas age is determined by weighting each age and associated uncertainty, based on the amount of ^{39}Ar$_K$ present, in each increment and averaging.

Method

Samples were crushed and sieved in order to isolate size fractions that contained the mica or amphibole as individual rather than composite grains. Standard density, magnetic and hand-picking techniques were used to purify mineral separates. In all cases, purity was estimated to be greater than 99 percent. The samples were encapsulated in aluminium foil,

Table 1: *(Concluded)*

Release Temp C	$(^{40}Ar/^{39}Ar)^*$	$^{37}Ar/^{39}Ar$	$(^{36}Ar/^{39}Ar)^*$	Moles ^{39}Ar	^{39}Ar % of Total	$\%^{40}Ar$ Radiogenic	K/Ca	Apparent Age(Ma)
				BIOTITE				
Sample #191; Pelite; Gander Group; J = .005989								
750	42.87	0.0107	0.0086	102.7	11.1	94.0	45.67	390.1 +/- 4.9
890	42.35	0.0054	0.0034	133.3	14.1	97.6	91.04	398.9 +/- 4.1
925	41.91	0.0089	0.0027	109.2	11.8	98.0	55.21	396.8 +/- 4.2
950	41.96	0.0206	0.0022	87.8	9.5	98.4	23.80	398.5 +/- 4.2
1025	42.01	0.0159	0.0022	53.9	5.8	98.4	30.76	398.9 +/- 4.0
1130	41.55	0.0131	0.0019	229.5	24.8	98.6	37.52	395.8 +/- 4.1
Fusion	41.39	0.0158	0.0011	209.3	22.6	99.1	30.97	396.5 +/- 3.9
Total				925.6	100.0		Total Gas Age	396.3 +/- 4.2
Sample #204; Pelite; Gander Group; J = .006053								
750	32.17	0.0684	0.0148	61.8	8.2	86.3	7.16	280.2 +/- 3.1
825	40.88	0.0216	0.0052	106.6	14.1	96.1	22.70	385.0 +/- 3.9
900	42.02	0.0160	0.0038	75.4	10.0	97.3	30.55	398.8 +/- 3.9
975	44.35	0.0265	0.0026	72.8	9.6	98.2	18.50	422.0 +/- 4.2
1050	45.14	0.0281	0.0020	152.1	20.1	98.6	17.44	430.4 +/- 4.3
1130	43.68	0.0312	0.0020	139.1	18.4	98.5	15.71	417.8 +/- 5.3
Fusion	41.08	0.0523	0.0023	147.6	19.5	98.3	9.37	394.5 +/- 4.1
Total				755.3	100.0		Total Gas Age	398.4 +/- 4.2
Sample #846; Deadmans Bay Granite; J = .006196								
750	37.46	0.0466	0.0094	80.0	6.9	92.5	10.51	351.0 +/- 3.7
825	40.77	0.0158	0.0023	146.3	12.6	98.2	31.04	399.8 +/- 4.4
900	41.36	0.0120	0.0025	175.5	15.1	98.1	40.94	404.8 +/- 4.1
975	42.04	0.0279	0.0020	118.3	10.2	98.5	17.58	412.2 +/- 4.0
1050	42.56	0.0327	0.0014	188.5	16.2	98.9	15.00	418.3 +/- 4.1
1130	41.41	0.0324	0.0014	269.8	23.2	98.9	15.11	408.2 +/- 4.1
Fusion	41.36	0.0566	0.0017	184.6	15.9	98.7	8.66	406.9 +/- 4.0
Total				1162.9	100.0		Total Gas Age	404.5 +/- 4.1

weighed and loaded into silica glass vials, which were then sealed. Samples were irradiated in the H5 facility of the Phoenix Memorial Laboratory at the University of Michigan. Correction factors for Ca and K-derived isotopes were determined through the analysis of irradiated K and Ca salts and are as follows: $(^{40}Ar/^{39}Ar)_K = 0.0347$; $(^{36}Ar/^{37}Ar)_{Ca} = 0.0002572$; $(^{39}Ar/^{37}Ar)_{Ca} = 0.0007494$. The primary flux monitor was MMhb-1 (Alexander *et al.*, 1978).

The irradiated samples were heated in a molybdenum crucible within an ultra-high vacuum extraction system using R induction heating. Extraction temperatures are estimated to be accurate to ± 50°C. Standard gettering techniques were used to purify the inert gases. The isotopic composition of Ar was determined using a 15 cm Nier type mass spectrometer. Ages were calculated using the equations of Dalrymple *et al.* (1981).

Results

The muscovite and biotite cooling ages discussed here, are for those samples collected in the Ocean Pond area only.

Muscovite. $^{40}Ar/^{39}Ar$ age spectra were determined on muscovite separates from 3 samples from the biotite zone, (pelite, samples 86-191, 86-204 and 86-253) and one sillimanite zone hornfels (sample 87-839), in the Jonathans Pond formation, Gander Group. All have well defined plateaus after initial increments that suggest some loss of Ar (not necessarily through natural means), but which is less than 10 percent of total Ar liberated. Samples 86-191 and 86-253 have total gas ages of 391.6 ± 6.8 and 390.17 ± 4.13 Ma respectively, and plateau ages of 394 ± 6 and 391 ± 4 Ma respectively. However, although sample 86-204 has a reasonable plateau, it is approximately 20 Ma younger than the others (and is also the youngest dated rock). The muscovite in this sample is very fine grained, < 0.1 mm, defines the main fabric and is strongly crenulated by F_3. Although the very fine grain size in the sample 86-204 may have effectively lowered the closure temperature, sample 86-253 has muscovite of similar grain size, but its age is similar to that defined by other samples. Also, the late microfolding is not as well developed in sample 86-253.

Sample 87-839 is a sillimanite-bearing semipelitic rock, approximately 4 km northeast of the Deadman's Bay Granite. The rock is a hornfels and comprises granoblastic quartz, plagioclase, muscovite and sillimanite needles, preserved in quartz. A fine grained equigranular, muscovite, biotite granite, which is slightly foliated, intrudes the semipelite. The muscovite is coarse grained and is located within the thermal aureole of the Deadman's Bay Granite. The age spectrum for sample 87-839 has a plateau age of 397 ± 5 Ma and a total gas age of 396.4 ± 4.6 Ma.

Samples 87-847 and 86-302 represent the Ocean Pond granite and a muscovite-rich pegmatite, spatially and mineralogically related to the granite. The age spectrum for the granite (87-847) is quite flat with a plateau age of 398 ± 5 Ma and a total gas age of 397.1 ± 4.7 Ma. The pegmatite sample (87-302) was collected 3 km southwest of the granite, intrudes the garnet zone and has a slightly younger age of 389 ± 4 (total gas age) and a plateau age of 390 ± 7 Ma.

Biotite. $^{40}Ar/^{39}Ar$ age spectra were determined for two biotite concentrates from pelitic samples 86-191 and 86-204, collected from the biotite zone of the Gander Group Jonathans Pond formation and one concentrate, sample 87-846, from the Deadman's Bay Granite; age spectra are shown on Figure 1.

The spectrum for sample 86-191, already mentioned, has a plateau age of 398 ± 5 and a total gas age of 396.3 ± 4.2 Ma.

Sample 86-204 has a discordant hump-shaped spectrum and gives a total gas age of 398 ± 4 Ma. The low temperature part of the spectrum records an anomalously young age, however, the initial gas fraction represents only 8 percent of the total gas liberated and thus the young age has little, if any, significance. The total gas age is similar to both plateau and total gas ages (biotite) of sample 86-191 and is also similar to plateau ages given by muscovite spectra of pelites from the same area. Sample 87-846, collected from the Deadman's Bay Granite, is a K-feldspar rich, megacrystic biotite granite, which contains 15 to 20 modal percent biotite and < 5 percent muscovite. The age spectrum is hump shaped and the first gas fraction represents 7 percent of the total Ar liberated. The remaining increments are slightly discordant and a plateau is absent; the total gas age is 404 ± 4 Ma.

DISCUSSION

$^{40}Ar/^{39}Ar$ incremental release spectra on muscovite and biotite from metasedimentary and igneous rocks from the northeastern part of the Weir's Pond map area, generally exhibit reasonable plateaus, and when combined with microfabric and porphyroblastesis relationships, suggest that a single regional thermal event affected most of these rocks.

Most muscovite plateau ages fall in the range between 404 and 388 Ma and are similar to the total gas ages, which range from approximately 400 to 385 Ma. Muscovite cooling ages on the Ocean Pond granite and an associated pegmatite, fall within this range. It is not clear whether the granite and metamorphism are two separate effects of a crustal heating event or whether the metamorphism reflects a regional aureole associated with a massive granite at depth; the Ocean Pond granite being a surface manifestation of this larger granitic body. Lux *et al.* (1986) in a study of New England granites and low pressure–high temperature metamorphic belts, concluded that the plutons provided the source of heat for the regional metamorphism. The relationship between the Ocean Pond granite and the regional metamorphism in the Weir's Pond area is very similar to that displayed by granites and metamorphic belts in New England.

Biotite cooling ages are similar to those of muscovite, and range from 402 to 393 Ma, implying that both minerals passed through their closure temperatures, of between 350°C and 250°C respectively.

Muscovite and biotite $^{40}Ar/^{39}Ar$ cooling ages suggest that these metasediments and the Ocean Pond granite cooled through greenschist facies conditions between approximately 400 and 385 Ma. The partial overlap of biotite growth and F_3 development, suggest that these cooling ages constrain the D_3 event, which was a phase of upright folding.

Although the Deadman's Bay Granite is clearly posttectonic, cutting all structures in the metasediments, its biotite age suggests that it cooled slightly earlier than the metasediments and the synkinematic Ocean Pond granite. The similarity between this biotite age and the coarse muscovite from the hornfelsed aureole, and the remaining muscovite and biotite from the Ocean Pond area and southwestward, suggest that the Deadman's Bay Granite intruded very soon after the main thermal event. As reference to Figure 1 will show, some of the muscovite and biotite samples were taken up to 20 km from the Deadman's Bay Granite and thus have not been affected by any localized thermal resetting due to this granite. The 404 Ma biotite age on the Deadman's Bay Granite is similar to a K/Ar age of 400 ± 13 Ma, obtained from the granite by Currie (*in* Stevens *et al.*, 1981).

Sericitic metavolcanic rocks from the western part of the Avalon Zone, yielded whole rock $^{40}Ar/^{39}Ar$ spectra with plateau ages in the range 370 to 390 Ma (Dallmeyer *et al.*, 1983), which were interpreted to closely date the regional metamorphism and associated deformation in the western Avalon Zone. ^{40}Ar/^{39}Ar age spectra on biotite from the Hare Bay Gneiss in the northeastern Gander Zone, record plateau ages ranging from 365 to 383 Ma (Dallmeyer *et al.*, 1981) and were interpreted to date the regional post-metamorphic cooling and also reflect a minimum date for initial tectonic activity along the Dover Fault. The older ages within the range of dates described by Dallmeyer *et al.* (1983) are similar to those presented in this paper, and imply a synchroneity in post-metamorphic cooling throughout the northern Gander and western Avalon zones.

In southern Newfoundland, Dunning *et al.* (1988) argue for a major Late Silurian orogeny with medium to high grade, regional metamorphism and deformation, combined with plutonism and volcanism. The North Bay Granite is posttectonic, dated by U–Pb at 396^{+5}_{-3} and provides a younger age limit for the orogenic event in the south. Cooling ages on biotite and muscovite from the Weir's Pond area in the northeastern Gander Zone, are approximately 20 Ma younger than U–Pb peak metamorphic ages for the southern Newfoundland event and suggest that plutonism and metamorphism in the northeastern Gander Zone may have occurred synchronously with the Late Silurian orogeny as defined for southern Newfoundland.

The overlap between biotite porphyroblastesis (and recrystallization of muscovite) and the D_3 deformational phase, allows this phase to be somewhat constrained to a Late Silurian–Early Devonian time frame. The early, main phase of deformation, which had quite a different style and was pre-metamorphic, is therefore an older event, probably Early Silurian at the youngest, based on correlations with southern Newfoundland.

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GEOLOGY OF THE CUTWELL GROUP, NOTRE DAME BAY, NEWFOUNDLAND: A DISRUPTED VOLCANIC–VOLCANICLASTIC ISLAND-ARC COMPLEX

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ABSTRACT

The rocks of the Cutwell Group record a three-stage structural history. The first deformation (D_1) *is Llanvirn or earlier, and resulted in northeast-directed thrusting and folding related to the sinistral strike-slip on the Long Tickle Fault. The second deformation* (D_2)*, of presumed Late Ordovician age, resulted in northwest-directed thrusting in response to a sinistral oblique-slip on the Long Tickle Fault. The third deformation* (D_3)*, which also includes rocks of the Silurian Springdale Group, resulted in a dextral strike-slip faulting on the Lobster Cove and Crescent Lake–Tommy's Arm faults.*

The rocks of the Cutwell Group are divided into three lithofacies: the tuff lithofacies, the lapilli tuff–tuff breccia lithofacies, and the primary volcanic lithofacies. The observed facies variations are consistent with the formation and evolution of the group in an island-arc setting.

Volcanogenic sulphide mineralization and alteration are related to pulses of felsic volcanism. Correlation of the mineralized felsic sequences with the apparently barren felsic rocks elsewhere in the group, suggests that the latter should have similar mineral potential.

INTRODUCTION

Detailed field studies of the Cutwell Group in the Notre Dame Bay area, north-central Newfoundland, have been carried out as part of the project, 'Metallogeny of the Buchans–Roberts Arm Belt'. The object is to determine the metallogeny of non-ophiolitic rocks in the western Dunnage Zone, and this report summarizes the results of the second field season.

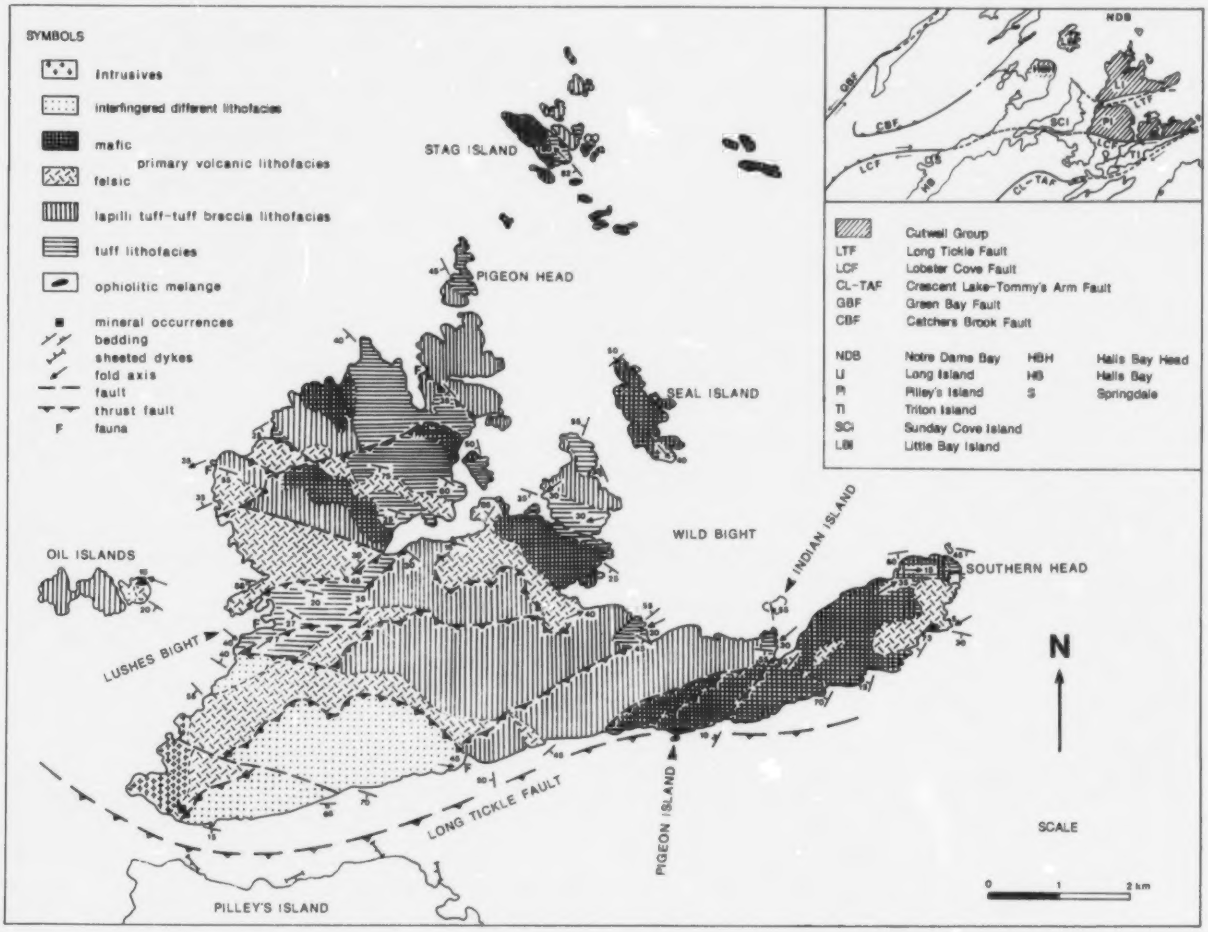
The Cutwell Group was defined by Kean (1973), based upon the detailed mapping of Long Island (Figure 1). He divided the Cutwell Group into seven stratigraphic units: six formations and one igneous complex, and interpreted the units as forming a continuous, over 5-km-thick, south-facing sequence. In addition, based on the chemical composition of volcanic rocks and on the observed stratigraphy, the Cutwell Group was inferred to represent an island-arc volcanic sequence, which recorded the transition from an immature to mature arc (Kean, 1973; Kean and Strong, 1975). Further, on the basis of the fauna from thin limestone lenses within volcanic rocks on the west shore of Little Bay Island and along the southern shore of Long Island, the Cutwell Group was tentatively assigned to the Middle Ordovician (Williams, 1962; Strong and Kean, 1972).

A Caradocian, fossil-based age for the black shale from the middle of the Cutwell Group was obtained by Dean (1977, 1978). He included the formation underlying the black shale in the pre-Caradocian 'immature island-arc' sequences, and the formations overlying the shale, in post-Caradocian 'mature island-arc' sequences. The deduced post-Caradocian age of the rock sequences in the southern part of the island was contradicted by an upper Arenig–lower Llanvirn age, indicated by the shelly fauna and conodonts from limestone beds on the southern shore of the island (Stouge, 1980). These conflicting ages suggest that there are major structural problems with the Cutwell Group, and these contradictions have to be resolved before a firm stratigraphy can be established.

Szybinski (1988) re-mapped the Cutwell Group and showed that it has been structurally disrupted. The structural style of the group was shown to be a relatively complicated thrust system, and the previously developed stratigraphic succession represents a series of structurally repeated lithological units. During the 1988 field season, survey work was concentrated on Long Island with the following objectives: 1) consolidating the stratigraphical and structural interpretations, 2) comprehensively documenting the structural boundaries between the units, and 3) to determine the correlation of the mineralized sequences with rocks elsewhere in the group.

REGIONAL GEOLOGY

The Cutwell Group lies within the Notre Dame Subzone of the Dunnage Zone (Williams *et al.*, 1988), which records the Cambrian and Ordovician history of the oceanic realm, commonly referred to as Iapetus. Rocks of the Cutwell Group


SYMBOLS
intrusives
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bedding
sheeted dykes
fold axis
fault
thrust fault
F fauna
STAG ISLAND
PIGEON HEAD
SEAL ISLAND
WILD BIGHT
INDIAN ISLAND
SOUTHERN HEAD
OIL ISLANDS
LUSHES BIGHT
LONG TICKLE FAULT
PIGEON ISLAND
PILLEY'S ISLAND
N
SCALE
0
1
2 km
NDB
GBF
CBF
LCF
SCI
LI
LTF
PI
TI
HB
CL-TAF
Cutwell Group
LTF Long Tickle Fault
LCF Lobster Cove Fault
CL-TAF Crescent Lake-Tommy's Arm Fault
GBF Green Bay Fault
CBF Catchers Brook Fault
NDB Notre Dame Bay
LI Long Island
PI Pilley's Island
TI Triton Island
SCI Sunday Cove Island
LBI Little Bay Island
HBH Halls Bay Head
HB Halls Bay
S Springdale

Figure 1. *Generalized geology of the Cutwell Group on Long Island.*

outcrop in western Notre Dame Bay, from the western shores of Halls Bay Head to the eastern end of Triton Island (Figure 1). Contacts with adjacent groups are faulted and two faults are of regional importance. The Lobster Cove Fault, a large strike-slip fault system (Szybinski, 1988; Calon and Szybinski, 1988) separates the Cutwell Group from the Lower Ordovician Roberts Arm Group on Pilley's and Triton islands. The Long Tickle Fault, an oblique slip fault, divides rocks of the Cutwell Group on Long Island, from sheeted dykes and pillow lavas of the Lushs Bight Group on Pilley's and Sunday Cove islands (Figure 1).

STRUCTURAL GEOLOGY

A preliminary compilation of the 1987 field data (Szybinski, 1988) showed that the volcanic rocks of the Cutwell Group were affected by two major deformational events; an older event, which resulted in extensive, north-directed thrusting and a younger event, which was recorded by major, strike-slip, dextral movement on the Lobster Cove Fault.

Re-examination of field data, deformation microstructures in oriented samples and an extensive kinematic study on Long Island and adjacent islets during the 1988 season, revealed that the deformational history of the area is more complex than originally thought and that the rocks of the Cutwell Group record at least two events of north-directed thrusting:

1) D_1 event, during which northeast-directed movement was associated with sinistral strike-slip on the Long Tickle Fault; and

2) the D_2 event, a northwest-directed thrusting and high-angle reverse faulting related to sinistral oblique slip on the Long Tickle Fault, which utilized some of the previously developed structures.

The latest deformational event, D_3, is recorded by dextral strike-slip on the Lobster Cove and Crescent Lake–Tommy's Arm faults, and may be related to movements on the Green Bay Fault.

The D_1 Event

The northeast sense of movement obtained from some kinematic indicators (e.g., orientation of fold hinges and fold vergence, shear bands and slickensides), in the central and northern part of Long Island may be explained by and is compatible with, sinistral strike-slip movement on the Long Tickle Fault. The Long Tickle Fault divides Long Island from Pilley's Island and the trace of this fault is mainly offshore.

Several folds within rocks of the Cutwell Group exhibit a 'soft-sediment' style of deformation, e.g., within the shale–limestone breccia–greywacke broken unit, on the southern shore of Lushes Bight Harbour or in tuffaceous rocks of the Wild Bight area (Plate 1). The relative consistency of the fold orientations with respect to orientation of cleavage in the immediate area, suggests that the folds may be valid kinematic indicators (Agar, 1988), and are assigned to the D_1 event. Northeast-directed movement is also deduced from deformation within the shaly mélange on Seal Island.

Plate 1. *'Soft-sediment' style of folding of mafic tuffs, Wild Bight area (looking southeast).*

Mafic dykes, acting as feeders to andesitic flows, cut through already deformed strata in the Wild Bight area. The andesitic flows that are laterally extensive, overlie shales and argillites of upper Arenig age (S.H. Williams, personal communication, 1988) on Southern Head. This and the presence of 'soft sediment' style folds in lower Llanvirn strata, allows the possibility that the age of first deformation may be Arenig–lower Llanvirn.

Locally, well developed S_1 axial-planar cleavage is overprinted by regional S_2 cleavage (Plate 2), and is related to the D_2 event.

The D_2 Event

The Long Tickle Fault on Long Island brings rocks of ultramafic basement(?) against rocks of the Cutwell Group. Ultramafic rocks are exposed on the southern shore of Long Island and on Pigeon Island, a small offshore island (Figure 1) and are predominantly metasomatic magnesite-rich mylonites. These mylonites and an ophiolitic mélange form a small-scale strike-slip exotic duplex (Woodcock and Fischer, 1986), which remains attached to the footwall. Mylonites within this duplex, exhibit locally well developed shear bands (C-surfaces; Berthé *et al.*, 1979), the geometry of which suggests development in sinistral oblique slip.

Plate 2. *F_1 fold with well developed S_1 axial-planar cleavage overprinted by S_2 cleavage, Wild Bight area (looking east).*

Structures compatible with sinistral oblique-slip are found immediately north of the Long Tickle Fault and across Long Island. The intensity of deformation related to this event (D_2), decreases toward the north. The D_2 deformational event is responsible for the present distribution of the lithostratigraphical units (Figure 1).

The D_3 Event

The geometry and sense of movement on the Lobster Cover Fault clearly show it to be a complex structure probably related to movements on the Green Bay Fault. The fault has commonly been interpreted on the basis of regional geology, as a folded, south-directed thrust fault (Dean and Strong, 1977). Contrary to this interpretation, deformation structures in Silurian red sandstones along the lineament on Sunday, Pilley's and Triton islands, indicate that it is a dextral strike-slip fault (Szybinski, 1988; Calon and Szybinski, 1988). This structure appears to truncate (at a very low angle) an earlier thrust system and sinistral oblique-slip shear zones (D_2). The Lobster Cove Fault changes in its geometry and orientation along the length of the structure. In the Springdale area, it forms a large-scale bend and grades from strike-slip in the east to oblique-slip and finally to a southeast-directed thrust in the west. In the latter area, rocks of the Lower Ordovician Catchers Pond Group are brought upon volcanics of the Silurian Springdale Group (Jenner and Szybinski, 1987).

Recently, Blewett and Pickering (1988) stressed the importance of sinistral shear during the Acadian Orogeny in north-central Newfoundland. Their interpretations and conclusions were based on field observations in the area, immediately to the south of the Lobster Cove Fault. For this reason, structural observations have been extended south of the Lobster Cove Fault into the Roberts Arm Group and the shear zone of the Crescent Lake–Tommy's Arm Fault, which form the southern shore of Triton Island. Field studies revealed close similarities in the structural history of both faults. The vergence of well developed vertical folds along this lineament indicate clearly a dextral sense of shear. However, evidence for northeast-directed thrusting and sinistral strike-slip have been found within shear zones transected by both dextral strike-slip faults.

Age of the D_2 and D_3 Events

There is some evidence to establish the relative and absolute ages of the D_2 and D_3 events. The D_3 dextral strike-slip is the youngest deformational event in the area and is at least Silurian or younger, since it incorporates the red sandstones and conglomerates of the Silurian Springdale Group. The Lobster Cove Fault postdates the D_2 event because it truncates the shear zones related to this event on Triton, Pilley's and Sunday Cove islands.

The lower age limit of the D_2 event may be determined from the youngest strata involved in the thrusting, which is of middle Llanvirn age (O'Brien and Szybinski, 1988). The upper age limit on the D_2 thrusting, in the Cutwell Group, is considered to be Late Ordovician, since Silurian strata appear not to be involved.

However, it is possible that structures assigned to the D_2 and D_3 deformational events may represent one and the same event.

STRATIGRAPHY

The structural model developed for the Cutwell Group in this paper disagrees with the previous stratigraphy established and formalized by earlier workers (Kean, 1973; Kean and Strong, 1975; Dean, 1977, 1978). In particular, units representing the same stratigraphic level, reappear in the different thrust slices of the formalized stratigraphy (Szybinski, 1988). For this reason, the previously established formal stratigraphical nomenclature is abandoned, and until the Cutwell Group is redefined and a new stratigraphy erected, it is discussed in terms of its lithofacies.

Age Criteria

Correlation of units between different thrust slices have been facilitated by the presence of abundant identifiable microfossils, in particular, conodonts and graptolites. Preliminary identification of the collected conodonts (O'Brien and Szybinski, 1988; F. O'Brien, personal communication, 1988) and graptolites (Williams, 1988; S.H. Williams, personal communication, 1988), has been sufficient to establish the validity of the structural model. Assemblages of the fauna of similar lower to upper (?) Llanvirn age were obtained from several scattered outcrops across the island revealing repetition of the stratigraphy. Dunning and Krogh (1988) have obtained a U–Pb zircon age of 469^{+5}_{-3} Ma, for a felsic tuff overlying the sedimentary fossiliferous unit of lower Llanvirn age, on Oil Island, which confirmed the fauna-based age. Older fossiliferous strata of upper Arenig age (S.H. Williams, personal communication, 1988) have also been found in the area and may be traced from Southern Head up to Pigeon Head, north of Long Island. The evolution of the volcanic centre(s) forming the Cutwell Group on Long Island covers a time span of at least upper Arenig to upper (?) Llanvirn. Lateral correlation of the Cutwell Group with the Buchans Group is suggested by the age and the assemblages of conodonts present within the felsic sequences.

Lithofacies

The rocks of the Cutwell Group on Long Island are divided into three lithofacies (Busby-Spera, 1988): 1) the tuff lithofacies; 2) lapilli tuff–tuff breccia lithofacies; and 3) the primary volcanic lithofacies.

The Tuff Lithofacies. The tuff lithofacies consist of finely laminated, cherty black shales interbedded with green reworked tuff (turbidite) and grey cherts. Locally, thin beds of greenish-black feldspar–phyric tuff are present in the upper part of this sequence. Intercalated horizons of welded to nonwelded dacitic pyroclastic flows are a common occurrence that cut down into pre-existing lithologies, and contain a large amount of shale and chert rip-ups (Plate 3).

The Lapilli Tuff–Tuff Breccia Lithofacies. The lapilli tuff–tuff breccia lithofacies is often inter-stratified with rocks of the tuff lithofacies. It consists of a variety of volcanic fragments up to 0.8 and 1 m in diameter, in a fine- to medium-grained tuffaceous matrix. The most common clasts are of plagioclase-phyric and clinopyroxene-phyric mafic volcanic and subvolcanic rocks, as well as felsic lithologies. Highly silicified and epidotized fragments occur and may indicate strong synvolcanic (prior to deposition) alteration. Locally, isolated blocks or clusters of blocks of limestone are present within this lithofacies. The variety of fragments within the lapilli tuff–tuff breccia lithofacies indicates mixing of volcanic debris, and is interpreted as being deposited from debris flows. Some of the debris flows, scoured channels in underlying rocks of the tuff lithofacies and contain rip-up fragments of the tuff lithofacies.

The Primary Volcanic Lithofacies. The primary volcanic lithofacies is diverse and includes dacitic domes and related subvolcanic intrusives, felsic tuffs and tuff-breccia, and mafic, basaltic to andesitic, massive and pillowed flows and pillow breccias.

Dacitic volcanic rocks vary from flow and intrusive phases of banded dacite to structureless plagioclase-phyric tuff. Locally, these phases are difficult to distinguish from highly altered mafic rocks. Felsic rocks are usually light green to buff and when associated with sulphide mineralization, they are yellow to rusty.

Plate 3. *Partially welded dacitic pyroclastic flow showing shale and chert rip-ups, cutting down into strata on Southern Head.*

The Southern Head area appears to represent part of a felsic dome, overlying mafic lavas, breccia and hematitic tuff. Flows, oligomictic felsic tuff breccia and slump folds have been recognized within dacitic rocks immediately overlying the mafic basement. A dacitic intrusive phase of the dome intrudes the underlying mafic rocks. A large dacitic feeder dyke, is also exposed, and is associated with an intrusive breccia of dacitic clasts and argillite and shale fragments.

Dacitic rocks (flows and pyroclastics) in the central and western parts of Long Island are associated with subvolcanic intrusives. The lens-like shape of the outcrops suggests that they may too represent fragments of dome(s).

Basaltic to andesitic, massive to pillowed flows and pillow breccias are a volumetrically important component of this lithofacies (Plate 4). They vary from aphanitic and clinopyroxene-rich basalt to plagioclase-phyric andesite. Massive flows often grade vertically into pillowed flows. Interpillow material is usually green to red chert. On Stag and Seal islands, hyaloclastite and tuffaceous limestone is often found between pillows. The ratio of massive or pillowed flows to volcaniclastic rocks varies across the island; in the northeastern part of the island the ratio is high whereas in the southwest, it is low.

Plate 4. *Pillow-breccia and mafic sheet-flow (delineated) on southern shore of Long Island.*

Stag Island Breccia

Stag Island and several islands in the vicinity are partially underlain by breccia, consisting predominantly of angular clasts of coarse-grained gabbro, diabase and granodiorite. Often, the breccia interfingers with, or is overlain by, pillow basalt. Locally, this lithology is intruded by plugs of gabbro and diabase dykes.

The stratigraphic position of the Stag Island breccia has not as yet been established, nor has a similar lithology been found elsewhere on Long Island. However, there is an outcrop of mafic tuff with angular clasts of coarse-grained gabbro, on the southeastern shore of Little Bay Island. These rocks seem to be structurally above (stratigraphically below ?) the unit, which contains limestone breccia of upper Arenig–lower Llanvirn boundary age (F. O'Brien, personal communication, 1988). If the correlation of these units is valid, then the breccia, associated pillow basalt and intrusives, form the oldest unit on Long Island. They may form the arc (?back arc) basement on which the volcaniclastic and primary lithofacies were deposited, in agreement with the model proposed by Kean (1973).

MINERALIZATION

Volcanogenic Sulphides

Several volcanogenic sulphide occurrences are found in the Cutwell Group, however, only two occurrences, the Oil Island and Shamrock, are of economic interest. The lack of good exposure especially of the Shamrock prospect, makes it difficult to determine the origin of mineralization. There is, however, an excellent outcrop of sulphide (predominantly pyrite) mineralization associated with dacitic volcanic rocks in the Southern Head area. Dacitic flows and pyroclastics overlie highly altered (chloritization–seritization) tuff, interbedded with hematite and magnetite-rich siliceous 'exhalite'. Pyrite forms the matrix in a strongly altered zone of brecciation, which is present within the dacitic rocks. The alteration zone is cut by an unaltered mafic dyke similar to the underlying pillow lava. This would suggest that both, mineralization and alteration, may be related to the exhalative processes associated with pulses of felsic volcanism. The presence of significant 'footwall' alteration along the northeastern shore of Long Island and the presence of mineralized welded dacitic pyroclastics within the shale-argillite sequence, together with the presence of a hematitic, siliceous 'exhalite' horizon, suggest an environment favourable for the deposition of massive sulphides.

Mineralized Veins

Several extensional quartz and calcite veins that have associated mineralization have been mapped in the Wild Bight area. These veins appear to be 'postkinematic', and are not related to Ordovician volcanism. Two kinds of veins predominate: 1) calcite veins with pyrite, sphalerite and galena, and 2) quartz veins with chalcopyrite, pyrite, galena, sphalerite and secondary malachite.

The veins are scattered over the entire area, but there are no larger scale mineral occurrences associated with them; they may be related to the Silurian or Devonian intrusive events.

Importance of Structural and Lithofacies Studies

Most volcanogenic massive sulphide deposits (VMS) in central Newfoundland are geograhpically associated with felsic volcanic rocks, and, like most VMS, are stratabound. In general, the mineralization is related to felsic domes and pyroclastic rocks, which occur in a restricted stratigraphic interval. However, structural complexity often obscures this simple relationship, e.g., in the Buchans camp.

In the Cutwell Group, there are several exploration targets that have characteristics found in VMS deposits. Both the structural and lithofacies analysis carried out in this study can be used to delineate the most favourable horizons for exploration. The structural model, here proposed for the Cutwell Group, is clearly important for any exploration program, since it is not possible to extrapolate the location of any favourable horizon using a layer-cake stratigraphic model.

SUMMARY

Field work resulted in new insights into the structural development of the western Notre Dame Bay area, and an informally revised stratigraphic succession of the Cutwell Group.

The conclusions based on this work include the recognition of three deformational events related to generally continuous sinistral transpression in the area. This resulted in two stages (D_1 and D_2 events) of the generally north-directed thrusting on Long Island associated with sinistral strike- and oblique-slip on the Long Tickle Fault. A final, late D_3 event resulted in dextral strike-slip on the Lobster Cove and Crescent Lake–Tommy's Arm faults.

The rocks of the Cutwell Group on Long Island are divided into three lithofacies:

1) the tuff lithofacies, comprising tuffaceous cherts, fine to sandy textured, reworked tuffs, locally interbedded with pyroclastic, partially welded flows,

2) the lapilli tuff–tuff breccia lithofacies comprising a variety of volcanic fragments in a fine- to medium-grained tuffaceous matrix, interpreted as deposited from debris and pyroclastic flows, and

3) the primary volcanic lithofacies comprising dacite domes, pyroclastics and intrusive phases, basaltic to andesitic massive and pillowed flows, and pillow breccias.

It has been recognized that volcanogenic sulphide mineralization and hydrothermal alteration on Long Island are associated with felsic volcanic rocks. These rocks, although only locally mineralized, seem to correlate with apparently barren felsic volcanics elsewhere in the stratigraphy. The already known occurrences, significant alteration and the presence of an hematitic 'exhalite' horizon, may be indicators that mineral deposits remain to be discovered on Long Island. This, and new information about the structure and informally revised stratigraphic succession, may provide the impetus for further exploration of the Cutwell Group.

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NEW GRAPTOLITE DISCOVERIES FROM THE ORDOVICIAN OF CENTRAL NEWFOUNDLAND

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ABSTRACT

Black shales on Long Island, western Notre Dame Bay that were previously assigned to the widespread 'Caradoc' black shale, actually belong to two stratigraphically distinct units associated with felsic volcanics and limestone breccia. The lower shale unit yields a late Arenig I. v. maximus *Zone fauna, whereas the upper unit contains a rich lower Llanvirn graptolite assemblage. The newly discovered graptolite faunas agree more closely with ages established using shelly macrofossils and conodonts, and are of open-ocean affinity, permitting an accurate correlation with many other localities both in western Newfoundland and throughout the world.*

Additional collecting and further studies of the graptolites from Snooks Arm on the Baie Verte Peninsula and Corner Pond at the southern end of Grand Lake, show the black shales to be lower Arenig (probably lower D. bifidus *Zone and* P. fruticosus *Zone, respectively) and close, but not identical in age. Thus, there is no one geographically widespread Lower Ordovician black shale in the Dunnage Zone, in contrast to that of the ubiquitous Middle Ordovician (Llandeilo-Ashgill?) unit.*

The Point Leamington Greywacke and greywackes of similar late Caradoc to Ashgill age in the Badger area, should probably be included within the Sansom Greywacke, with the base defined at the first coarse greywacke unit. However, formal definition is withheld pending further investigation. Some range in the timing of the onset of greywacke deposition throughout the area is evident, but this only encompasses an interval equivalent to parts of the D. clingani *and* P. linearis *zones of southern Scotland (perhaps two million years of less). Systematic change across the region cannot be demonstrated and previous claims for such a hypothesis included evidence from localities where the black shale is in tectonic, rather than stratigraphic, contact with greywacke. Graptolites have been recovered from an interval of alternating black and grey silty shales within the Point Leamington Greywacke at Point Leamington, indicating a correlation with the* D. anceps *Zone (mid to late Ashgill) of southern Scotland. These are by far the latest Ordovician graptolites discovered anywhere in Newfoundland, confirming the possible presence of the Ordovician–Silurian boundary within the Point Leamington Greywacke and Goldson Conglomerate.*

INTRODUCTION

This report summarizes the findings made during the past year, relating to the study of graptolites from central Newfoundland. Although the primary objective is a biostratigraphic and taxonomic revision of faunas from the widespread Middle Ordovician black shale unit (see Williams, 1988), the majority of collecting carried out during the summer of 1988 was from much earlier units, and from greywackes overlying the shale. Most of this work is directed at providing high-resolution correlations in order to better understand the geological evolution of the area. However, several localities have provided exciting and totally unexpected age dates, particularly on Long Island and at Point Leamington (Figure 1). Additional research will be undertaken on other Lower Ordovician sediments during the next two years, which will no doubt permit further refinement of the geological history.

LOWER ORDOVICIAN STRATA

Long Island

The stratigraphy on Long Island, northwestern Notre Dame Bay, is currently being revised by A. Szybinski (Department of Earth Sciences, Memorial University of Newfoundland), as part of his Ph.D program in the Halls Bay area. Included within the Cutwell Group is black siliceous shale, which previously has been assigned by Dean (1978, p. 121) and others, to the 'Caradoc' black shale of central Newfoundland. Dean (1978, p. 204) recorded four graptolite taxa from the south side of Lushes Bight harbour (Figure 2, locality 3), which had been identified by D. Skevington as:

Climacograptus sp. cf. *C. brevis* Elles and Wood
Cryptograptus tricornis (Carruthers)
Didymograptus sp. cf. *D. saggiticaulis* Gurley
Glyptograptus sp. cf. *G. euglyphus* (Lapworth)

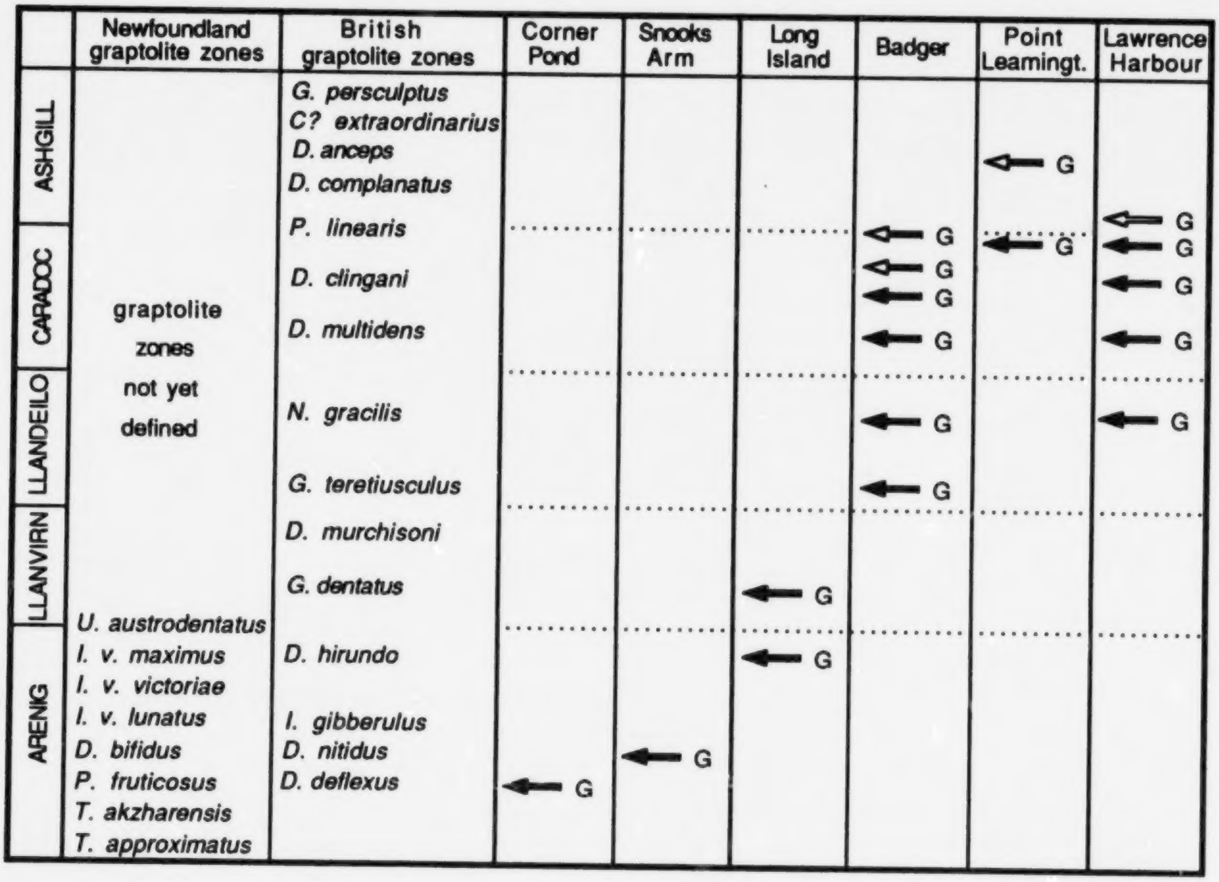

Figure 1. *Graptolite occurrences in the Ordovician of central Newfoundland discussed in the present paper; closed arrows indicate graptolites in black shale, open arrows in greywacke; additional data is taken from Williams (1982a, 1982b, 1988) and Williams and Stevens (1988).*

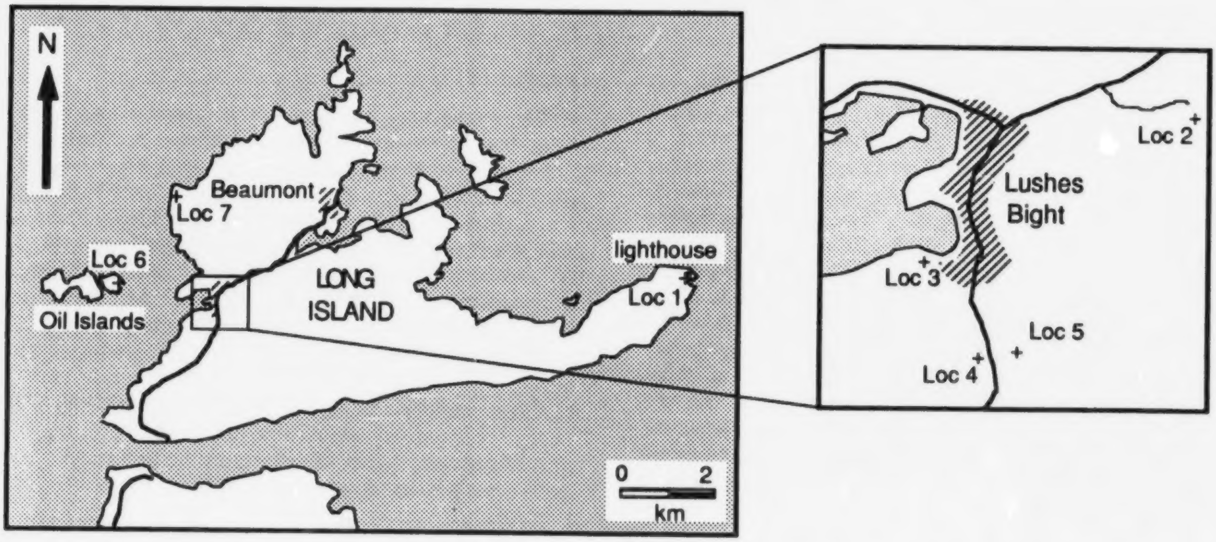
N
Beaumont
Loc 7
Loc 6
Oil Islands
LONG ISLAND
lighthouse
Loc 1
0
2
km
Loc 2
Lushes Bight
Loc 3
Loc 5
Loc 4

Figure 2. *Sketch map of Long Island, showing position of localities discussed in text.*

Dean (1978, p. 125) concluded that this assemblage represented a 'definite *Nemagraptus gracilis* Zone fauna'. As the shales were overlain by felsic volcanics, it was concluded that post-Caradoc volcanism had occurred in northern Notre Dame Bay, north of the Lobster Cove–Chanceport Fault, in contrast to the entirely pre-Caradoc Ordovician volcanism elsewhere to the south. However, the assigning of this unit to the 'Caradoc' black shale was in conflict with evidence from other faunal groups, particularly conodonts in limestones on the southern shore of Long Island, which were subsequently recorded as Llanvirn (Stouge, 1980). Strong and Kean (1972) had previously recorded a shelly macrofauna from the same limestone beds, which they considered to indicate a tentative 'Middle Ordovician' age. Elsewhere, conodonts were recorded by Williams (1962) from rocks on Limestone Island (offshore Little Bay Island), which are probably equivalent to those on Long Island, and the fauna identified as indicating 'early Middle Ordovician (Llanvirn?)'.

The two diplograptid species listed by Skevington (*in* Dean 1978, p. 204) are small, indistinct forms, while taxa similar in appearance to *C. tricornis* and *D. saggiticaulis* range from late Arenig to late Llandeilo–early Caradoc. The latter, combined with the apparently contradictory faunal evidence outlined above, questions the reliability of Skevington's graptolite identifications and subsequent correlation. New samples from Dean's original locality, and from two additional roadside outcrops in Lushes Bight (Figure 2, localities 4 and 5), failed to reveal more diagnostic material or identifiable taxa. However, the strata were only sparsely graptolitic, in marked contrast to the black shales belonging to the *N. gracilis* Zone elsewhere in central Newfoundland, casting further doubt on the age identification.

Szybinski (personal communication, 1988) subsequently discovered a few poorly preserved graptolites in a quarry, off the road leading from Lushes Bight to Beaumont (Figure 2, locality 2). Last summer (1988), additional material was collected from the quarry, which yielded a rich, abundant fauna. The shale unit is associated with the limestone breccia and overlain by felsic volcanics, demonstrating that it is the same unit exposed elsewhere in Lushes Bight. The collection has yet to be systematically described and identified, but provisional identification of the fauna includes the following:

Sinograptus typicalis Mu
Cardiograptus crawfordi Harris
Xiphograptus sp.
Acrograptus sp. cf. *A. cognatus* (Harris and Thomas)
Cryptograptus sp. cf. *C. antennarius* (J. Hall)
Paraglossograptus sp. cf. *P. tentaculatus* (J. Hall)
Skiagraptus sp. cf. *S. gnomonicus* Harris and Keble
Undulograptus? sp.
diplograptid spp.
dendroid sp. indet.

This assemblage, particularly the presence of *Sinograptus* sp., indicates an unequivocal lower Llanvirn age. It is similar to the fauna recorded by Lenz and Jackson (1986) from the *P. tentaculatus* Zone of the Canadian Cordillera. To a slightly lesser extent it compares to that listed by Finney and Skevington (1979), from the Table Cove Formation of the Table Head Group, western Newfoundland and considered by them to be equivalent to the *D. decoratus* Zone (Darriwilian 3) of Victoria, Australia. The same assemblage was also found at an additional locality on the coast north of Lushes Bight (Figure 2, locality 7), and the sparse remains found at the previous three localities can also be attributed to these taxa. Such a rich, cosmopolitan fauna is clearly of open-ocean type (see Fortey and Cocks, 1986), in direct contrast to that found in coeval black shales deposited in marginal, partially restricted areas, such as those of the type Llanvirn in South Wales.

An additional exposure of highly siliceous, black shale on the easternmost point of Long Island, about 200 m north of the lighthouse (Figure 2, locality 1) proved graptolitic and yielded the following forms:

Xiphograptus svalbardensis (Archer and Fortey) (deflexed form)
Isograptus victoriae maximus Harris
Pseudisograptus ensiformis (J. Hall)?
sigmagraptid sp.
dendroid sp. indet.

This assemblage indicates the late Arenig *I. v. maximus* Zone, defined recently by Williams and Stevens (1988) in the Cow Head Group of western Newfoundland, and shows the shale to be a distinctive stratigraphic unit from that of Lushes Bight. Such an assemblage is of open-ocean affinity (Fortey and Cocks, 1986). However, the shale is both underlain and channelled by welded tuffs, suggesting small, steep-sided, oxygen-starved basins adjacent to explosive volcanism.

The following conclusions regarding the black shales on Long Island can be drawn from the data presented: 1) Two distinct black shale units are present, one of late Arenig (*I. v. maximus* Zone) age and one belonging to the early Llanvirn; 2) both graptolites and other fauna (conodonts and shelly macrofossils) are in broad agreement concerning ages and correlation; 3) the intimate association of graptolitic sediments and felsic volcanics is potentially very promising for the linking of radiometric dating with graptolite biostratigraphy and the Arenig–Llanvirn boundary; 4) there was no post-Caradoc Ordovician volcanism on Long Island; it was entirely pre-Caradoc, as found elsewhere in central Newfoundland, and spanned a period of perhaps five million years from late Arenig to early Llanvirn; and 5) the graptolite faunas represent open-ocean assemblages, intimately associated with relatively near-shore volcanism (e.g. welded tuffs).

Snooks Arm

Graptolites were first recorded by Snelgrove (1931) from Snooks Arm on the east coast of the Baie Verte Peninsula. His specimens were identified by Ruedemann as *Loganograptus logani* (J. Hall) and *Didymograptus gracilis* (Törnquist) (also see Ruedemann, 1947, p. 59), indicating an Early Ordovician (Arenig) age. The area was subsequently remapped by Neale (1957) and Upadhyay (1973), with both authors quoting Snelgrove's Early Ordovician age. Upadhyay (1973) divided the Snooks Arm Group into several formations; his lithostratigraphy was employed by DeGrace *et al.* (1976), who erroneously recorded that Snelgrove's graptolites originated from the Bobby Cove Formation (DeGrace *et al.*, 1976, p. 10). The fauna actually originates from near the base of the Balsam Bud Cove formation, as recorded by Dean (1978, p. 86). Dean (1978), D. Skevington and others collected additional samples, which apparently substantiated the Arenig age. No additional taxonomic identifications have been given since Snelgrove's (1931) original work.

During the summer of 1988, Snelgrove's original locality on the southern shore of Snooks Arm, approximately 400 m southeast of the harbour, was relocated and new material collected (Figure 3, locality 1). Despite heavy fracturing and shearing within the argillites, a much more diverse fauna than previously recorded was obtained, including:

Tetragraptus bigsbyi (J. Hall)
Tetragraptus cf. *T. reclinatus* Elles and Wood
Xiphograptus svalbardensis (Archer and Fortey)
Phyllograptus typus J. Hall
Acrograptus gracilis (Törnquist)
Etagraptus harti (T.S. Hall)
Graptoloid gen. A (*of* Williams and Stevens, 1988)

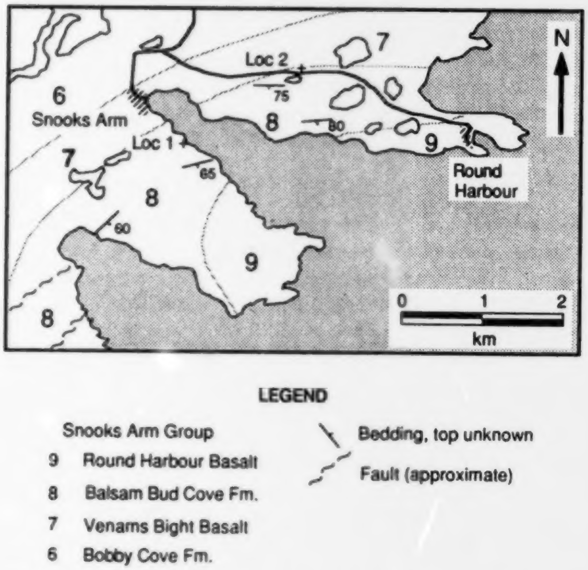

Figure 3. *Geological sketch map of Snooks Arm (after DeGrace et al., 1976), indicating position of graptolite localities.*

These taxa also occur in the Cow Head Group of western Newfoundland; the assemblage compares closely with that recorded by Williams and Stevens (1988) for the *D. bifidus* Zone of lower–middle Arenig age.

Several additional exposures of black cherty argillite occur beside the road between Snooks Arm and Round Harbour. Despite a well-developed, cross-cutting cleavage, one of them (Figure 3, locality 2) yields rare graptolites including *Pendeograptus fruticosus* (J. Hall). If, as is almost certain, these exposures belong to the same graptolitic unit seen in Snooks Arm, the presence of *P. fruticosus* further enhances the precision of correlation with western Newfoundland, as this assemblage is only found there in the lower half of the *D. bifidus* Zone.

Corner Pond

Within the Dunnage Zone, the only locality farther south that has been reported to yield Arenig graptolites is at Corner Pond, near the southeastern end of Grand Lake (Figures 4 and 5). The black shales and siliceous shales were mapped by Riley (1957) as part of the Glover formation; collections of graptolites were first made in 1970 by P. Dimmell (Noranda Limited) and F.D. Anderson (Geological Survey of Canada). Their collections were studied by R.B. Rickards, whose identifications were recorded by Dean (1976, p. 238) to include the following:

Tetragraptus fruticosus (J. Hall)
Tetragraptus pendens Elles
Tetragraptus cf. *T. reclinatus* Elles and Wood
Loganograptus logani (J. Hall)
Sigmagraptus sp.
Didymograptus extensus (J. Hall)
Didymograptus cf. *D. constrictus* (J. Hall)
Didymograptus cf. *D. similis* (J. Hall)

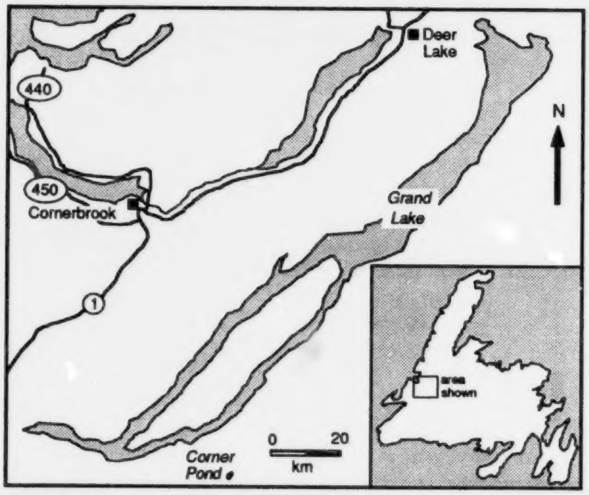

Figure 4. *Sketch map showing location of Corner Pond.*

Additional material was collected by H. Williams in 1977, which D. Skevington stated to include *Phyllograptus anna* J. Hall and *Didymograptus protobifidus* Elles, in addition to those forms reported by Rickards (*in* Williams and St-Julien,

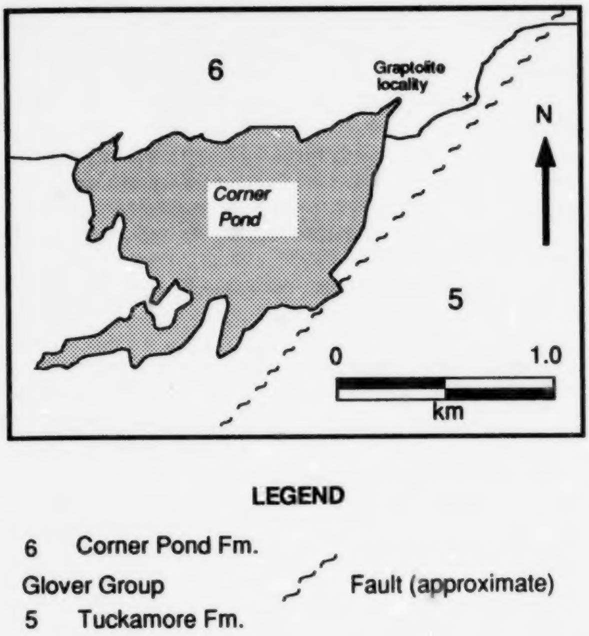

Figure 5. *Geological sketch map of Corner Pond (after Knapp, 1982), indicating site of graptolite locality.*

1978, p. 227). Both Rickards and Skevington considered the fauna to indicate the 'Middle Arenig *D. nitidus* Zone'; Williams and St-Julien (1978) further considered the age to indicate a correlation between the strata of Corner Pond and the Snooks Arm Group of Green Bay.

The locality was subsequently described by Knapp (1982, p. 251-255, Figure 8.8), who considered the black shale to belong to the Corner Pond formation rather than the Glover formation, from which it is separated by a fault. In addition to the black shale, the Corner Pond formation also contains red and purple shale, pillow lava, sandstone, conglomerate and a single limestone. Knapp (1982, p. 250) recorded a shelly macrofauna and conodonts from the latter unit, identified by G. Nowlan in 1981 as:

Drepanodus sp.
Periodon flabellum (Lindström)
Protopanderodus rectus (Lindström)
Scolopodus n. sp. cf. *S. gracilis* (Ethington and Clark)
Teridontus? sp.

Nowlan and Thurlow (1986, p. 60) summarized these findings suggesting a mid- to late Arenig age, and further stated that the graptolites indicated the 'mid-Arenig *D. protobifidus* Zone'.

The collection, made during 1970, is now housed in the paleontological collections of the Geological Survey of Canada, Ottawa. Examination of some of the material in light of the author's recent revision of Arenig graptolites in the Cow Head Group (Williams and Stevens, 1988), confirms most of Rickard's earlier identifications. Employing the taxonomic revisions made by Williams and Stevens (1988), the samples studied include:

Pendeograptus fruticosus (J. Hall)
Pendeograptus sp. cf. *P. pendens* Elles
Tetragraptus reclinatus Elles and Wood
Loganograptus logani (J. Hall)
Goniograptus thureaui (M'Coy)
Didymograptus (Expansograptus) extensus (J. Hall)
Didymograptus (Expansograptus) constrictus (J. Hall)
Didymograptus (Expansograptus) similis (J. Hall)

This assemblage could be found in both the *P. fruticosus* and *D. bifidus* zones of the Cow Head sequence, but the relative abundances of the faunal elements, the common occurrence of *D. (E.) extensus* and *P. fruticosus* specimens with four stipes, and the absence of *Xiphograptus svalbardensis* (Archer and Fortey) suggest a more likely correlation with the *P. fruticosus* Zone. Interestingly, *D. (E.) constrictus* is restricted to the previous *T. akzharensis* Zone in western Newfoundland, although it has been recorded from later intervals elsewhere (see Williams and Stevens, 1988, p. 48); no species of *Loganograptus* were found in the Cow Head Group. If the report of *Didymograptus (Didymograptellus) protobifidus* (probably *D. (D.) bifidus*) from this locality is correct, it suggests either that the assemblage actually belongs to the lower part of the *D. bifidus* Zone as seen at Snooks Arm or, more likely, that the graptolitic shale exposed at Corner Pond spans both the *P. fruticosus* and *D. bifidus* zones.

Conclusions

Lower to Middle Ordovician (Arenig to Llanvirn) graptolitic shales occur sporadically at several widely dispersed localities within the Dunnage Zone (Figure 1). Unlike the Middle Ordovician (Llandeilo to Ashgill?) black shale, none are widespread. Such localized occurrences of graptolitic black shale are atypical; most units throughout the world are laterally extensive and signify periods of high sea-level stand or widespread anoxia. Those discussed here clearly represent deposition in restricted, periodically anoxic basins, which have sufficient connections to normal oceanic environments to permit habitation by open-ocean graptolite faunas. Such an observation lends support to the currently popular model, whereby central Newfoundland is considered to have formed in a marginal plate setting with a complex series of back-arc and/or fore-arc basins. Additional graptolite research should permit further elucidation and unravelling of the complex geological history of this area.

UPPER ORDOVICIAN STRATA

Badger

Most graptolites previously recorded from the Badger area originate from the Middle Ordovician black shale unit, commonly referred to as the 'Caradocian black shale', and probably range in age from late Llandeilo to late Caradoc (informal zones 1 to 3 of Williams, 1988). Kean and Jayasinghe (1982) referred to this as Unit 4, which provided the material used by Erdtmann (1976) in his study of graptolites from the Exploits Group. Kean and Jayasinghe (1982, p. 18) also recorded graptolites from their overlying Unit 5 at one locality just east of Badger (this was inadvertantly incorporated in the graptolite assemblages of Unit 4 in Appendix A of their report). The graptolites from Unit 5 were identified by J. Riva (*in* Kean and Jayasinghe, 1982) as *Pseudoplegmatograptus* sp. (=*Arachniograptus* sp.) and *Climacograptus* cf. *tubuliferus* Lapworth, which he suggested indicated a late Caradoc or early Ashgill age (*P. linearis* or possibly *D. complanatus* Zone of the British sequence).

It should be noted that the graptolite locality is not as positioned on the published map (Kean and Jayasinghe, 1982), but occurs in a road-cut on the south side of the Trans-Canada Highway, 1.5 km southeast of the Badger turnoff (see Figure 6, locality 1). The section consists of an overturned, westerly dipping sequence of thinly bedded greywacke with subordinate interbedded black and grey shale. New collections from the shale have revealed the following forms:

Climacograptus tubuliferus Lapworth
C. miserabilis Elles and Wood
Orthograptus quadrimucronatus (J. Hall)
O. amplexicaulis (J. Hall) *sensu lato*
Glyptograptus daviesi Williams?
Plegmatograptus sp. cf. *P. nebula* Elles and Wood

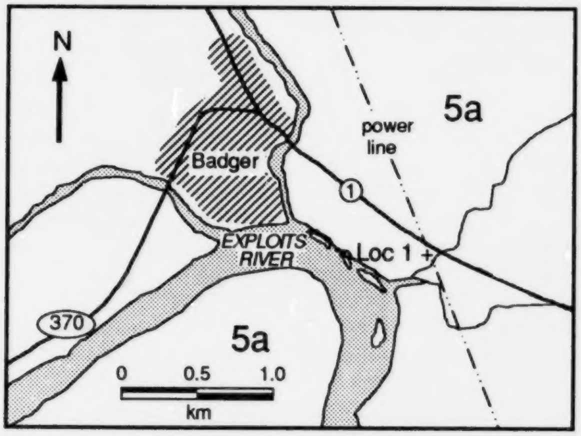

Figure 6. *Geological sketch map of Badger (after Kean and Jayasinghe, 1982) showing position of graptolite locality discussed in text.*

The assemblage is composed entirely of forms that also occur in the lower Hartfell Shale of southern Scotland (Williams, 1982a). Thus, it seems more likely to indicate a correlation with the British *P. linearis* Zone (late Caradoc or early Ashgill—see Williams and Bruton, 1983, for discussion) and informal zone 4 of Williams (1988), rather than the *D. complanatus* Zone (Williams, 1987), due to the occurrence of *O. quadrimucronatus* and possible presence of *G. daviesi*.

Kean and Jayasinghe (1982, p. 18) divided their Unit 5 into two; the lower subunit 5a (which yielded the graptolite fauna), consists of sandstone, minor conglomerate, shale and chert, whereas the upper subunit 5b is less extensive and consists of conglomerate. The authors correlated subunit 5a with the Sansom Greywacke of Notre Dame Bay and considered subunit 5b as a likely equivalent of the Goldson Formation. These suggestions appear eminently reasonable on both faunal and lithological grounds, and would further suggest that both subunit 5a and the Point Leamington Greywacke should be assigned to the Sansom Greywacke, which has historical precedence owing to its earlier definition by Heyl (1936).

The transition from continuous shale to alternating greywacke and shale, is exposed at Little Red Indian Falls, 19 km southwest of Badger (Williams, 1988). The graptolite fauna from this interval is dominated by *Climacograptus caudatus* Lapworth, which Williams (1982a, p. 252) recorded as typical of the lower part of the *D. clingani* Zone in southern Scotland. Therefore, deposition of clastics in this area, equivalent to the Sansom and Point Leamington greywackes, began in zone 3 of Williams (1988), and extended upward into zone 4. The thickness of succession separating the two graptolitic levels recorded here is unknown, as is the upper age limit of subunit 5a.

Point Leamington and Lawrence Harbour

Much has been informally discussed concerning the base of the Point Leamington Greywacke since it was first recorded as being diachronous by Bergström *et al.* (1974). The problems addressed here are twofold; first, how should the base of the Point Leamington Greywacke be defined and second, are the observations regarding age variation correct?

When Helwig (1967, 1969) defined the Point Leamington Greywacke, he included the upper section of the Middle Ordovician black shale within its base, subsequently referred to as 'an unnamed argillite' (Bergström *et al.*, 1974, p. 1634). Dean (1978, p. 135) considered these shales to be so distinct and important a lithostratigraphical marker that they should not be included within the formation. This author concurs with the view of Dean (1978); lithologically they are almost (if not entirely) inseparable from the Lawrence Harbour Shale. The distinction was based in most cases on purely faunal evidence (i.e., *'N. gracilis'* or *'D. multidens'* zones as opposed to *'D. clingani'* or *'P. linearis'* zones). A stratigraphic or regional structural break between the older and younger strata has always been assumed, although no real evidence has been presented for a hiatus. Formal redefinition of the various lithostratigraphical units will require additional research, but the present paper follows Dean (1978) in assuming the base of the Point Leamington Greywacke to be marked by the first recognizable greywacke bed. This means that the contact with the underlying unit is gradational, with a passage through increasingly coarser and paler grey silts and shales; many formation boundaries are similarly diffuse and no problems in subsequent definition are envisaged.

Bergström *et al.* (1974, p. 1635) stated that shales underlying the Point Leamington Greywacke at Cull Island, Mussel Bed Island, Western Arm Brook and at Cooper's Farm, Point Leamington, belong to the *D. clingani* Zone or to an interval transitional with the *P. linearis* Zone, whereas those at Lawrence Harbour are of *P. linearis* Zone age. They concluded from this evidence that the base of the greywacke was older to the west and south of New Bay than it was on the eastern side. This observation has been quoted widely in subsequent literature (e.g., Dean, 1978); however, considering that most graptolite zones in the Ordovician span one to two million years, and the diachronism encompasses perhaps half a zone, its significance is marginal at best. Furthermore, current investigations at both Cull Island and Western Arm Brook show quite clearly that the graptolitic shales at the first locality are not near the greywacke, whereas at the second locality they are in fault contact. Many additional localities remain to be visited, but to date the gradational interval between black shale and greywacke has been found at only one exposure in Notre Dame Bay, namely Lawrence Harbour (Figure 7, locality 2). Here shales associated with the transition yield a low diversity assemblage including:

Leptograptus flaccidus macer Elles and Wood
Dicellograptus pumilus Lapworth
Orthograptus amplexicaulis (J. Hall)

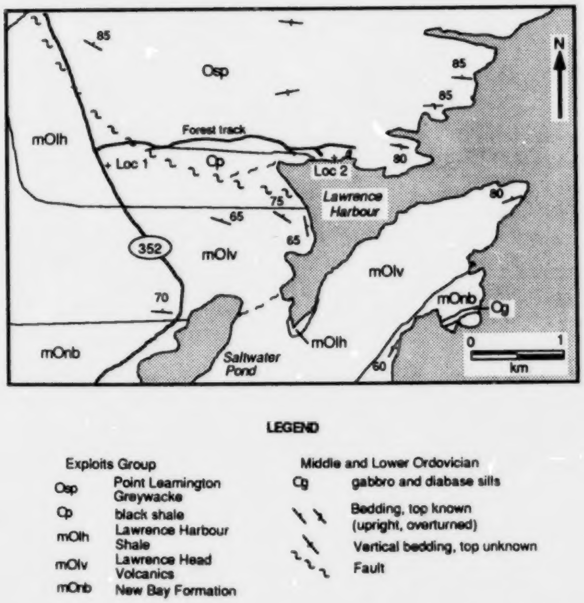
N
85
Osp
85
85
mOlh
Forest track
Loc 1
Cp
Loc 2
80
75
Lawrence Harbour
80
65
352
mOlv
65
mOlv
70
mOnb
Cg
mOnb
Saltwater Pond
mOlh
60
0
1
km
LEGEND
Exploits Group
Osp
Point Leamington Greywacke
Cp
black shale
mOlh
Lawrence Harbour Shale
mOlv
Lawrence Head Volcanics
mOnb
New Bay Formation
Middle and Lower Ordovician
Cg
gabbro and diabase sills
Bedding, top known (upright, overturned)
Vertical bedding, top unknown
Fault

Figure 7. *Geological sketch map of Lawrence Harbur (after Dean and Strong, 1975), indicating position of graptolite localities discussed in text.*

Possible specimens of *Pleurograptus linearis* (Carruthers) are also present, but confirmation of this taxon awaits further material. *D. pumilus* is characteristic of the *P. linearis* Zone in southern Scotland (Williams, 1982a), so despite the paucity of material, it is almost certain that the base of the greywacke at Lawrence Harbour falls within an interval equivalent to the *P. linearis* Zone of Britain (informal zone 4 of Williams, 1988).

Findings from this study thus confirm a diachronous onset of greywacke deposition following the Middle Ordovician black shales (Figure 1), but over a wider area than that discussed by Bergström *et al.* (1974). Until additional localities are found with complete transitions between the shale and greywacke, and the Point Leamington Greywacke is formally recognized as being either distinct or synonymous with the Sansom Greywacke (see above), no conclusions regarding systematic variation in age can be made.

The locality north of Point Leamington recorded by Bergström *et al.* (1974, table 4) as yielding a *P. linearis* Zone fauna was relocated during the present study (Figure 8, locality 3), but proved to be a fault-bounded, highly deformed exposure of the underlying black shale. The low diversity faunal assemblage, dominated by *Orthograptus quadrimucronatus* (J. Hall) could equally well be found in the *D. clingani* Zone and is here considered biostratigraphically inconclusive. Two new graptolite localities, within a sequence of alternating grey and black silty shales, were discovered last summer (1988) much nearer Point Leamington (Figure 8, localities 1 and 2). These yielded locally abundant specimens of *Climacograptus longispinus*

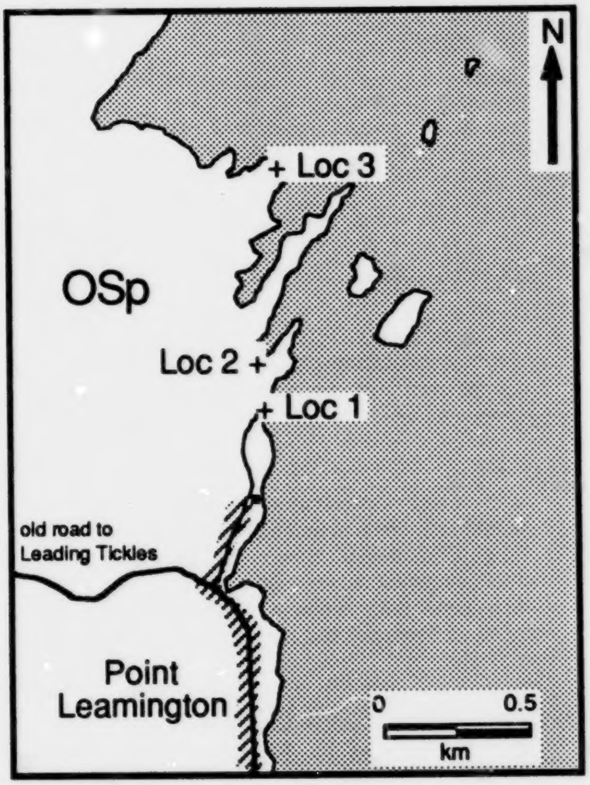
N
+ Loc 3
OSp
Loc 2 +
+ Loc 1
old road to Leading Tickles
Point Leamington
0
0.5
km

Figure 8. *Geological sketch map of Point Leamington (after Dean and Strong, 1975), indicating position of graptolite localities discussed in text (Osp = Point Leamington Greywacke).*

supernus Elles and Wood and *Orthograptus abbreviatus* Elles and Wood, both characteristic of the *D. anceps* Zone in southern Scotland (Williams, 1982b). The stratigraphical position of this very distinctive facies within the Point Leamington Greywacke has yet to be established. However, it demonstrates unequivocally that greywacke deposition continued into at least middle–late Ashgill times. This was assumed previously, but was unproven on faunal grounds, with the exception of supposed Llandovery corals in limestone blocks within an olistostrome near the gradational boundary with the overlying Goldson Formation (Helwig, 1967, 1969; Pickering, 1987).

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Geological Survey of Newfoundland, Report 89-1, pages 159-166*

CHEMICAL AND PHYSICAL PROPERTIES OF THE CANADA HARBOUR MARBLE DEPOSIT

A.F. Howse
Mineral Deposits Section

ABSTRACT

The Canada Harbour white marble prospect is the southernmost deposit in a continuous belt, which has been documented to extend from Canada Bay to Croque, a distance of more than 50 km. The marble occurs along deformed zones of Lower Ordovician carbonate rocks that structurally underlie the Hare Bay Allochthon.

Chemical analyses of chip samples, systematically collected from the face of the main open cut at Canada Harbour, show that about one fifth of the white marble is dolomitic. This feature distinguishes the deposit from the calcitic marble at the Penny's Pond and Coles Pond deposits. Silica, the chief contaminant is significantly higher in the Canada Harbour marble. This reflects the elevated amounts of impurities, especially quartz, associated with the dolomite units. Total iron as Fe_2O_3, and Al_2O_3 average less than one half percent in the deposit.

The average dry brightness of the Canada Harbour marble (95.96 percent light reflectance) is similar to that of the Penny's Pond (95.83 percent) and Coles Pond (95.95 percent) marble.

Preliminary comparisons indicate the Canada Harbour marble may closely resemble marble deposits identified in the White Arm window, near Croque.

INTRODUCTION

The first determined effort to exploit Newfoundland's marble resource took place at Canada Harbour early in this century. During the period 1912-15, William Edgar, a native of Scotland, tried to quarry white marble for dimension stone, but after a considerable investment of men and equipment, the enterprise failed. The inflationary freight rates and shipping hazards of World War I are believed to have been significant factors, which contributed to its failure (Martin, 1983).

The commercial potential of using Canada Harbour marble for building stone, ornamental stone and in memorials continued to attract interest after the First World War. Initial work by Muir (1935) and Howse (1936) on marbles in the White Bay and Canada Bay areas led to a more comprehensive study by Bain (1937), on its potential use for dimension stone. In 1967, British Newfoundland Exploration Limited investigated the Canada Harbour deposit in a program that included detailed geological mapping and bulk sampling (Barron, 1967), but nothing further developed. A survey of Newfoundland's limestone resources by DeGrace (1974) included an evaluation of the Canada Harbour deposit. This study was primarily concerned with identifying limestone and marble having the chemical specifications for such industrial applications as cement and quicklime manufacturing.

In 1985, the Newfoundland Department of Mines began a reassessment of the province's marble resource. The study was sparked by the increasing industrial demand for ground calcium carbonate for use as mineral filler in a variety of industrial products including paint, paper, and plastic. The 1985 survey involved extensive sampling of white, high-purity marble prospects in western Newfoundland (Howse, 1986). In order to meet the requirements of a superior grade filler, marble must have a high-calcium content and possess a minimum of acid-insoluble components such as quartz and other abrasive minerals. High brightness, in the order of 94 to 96 percent light reflectance (compared to a pure ASTM standard), is essential for a premium-grade product (Guillet and Kriens, 1984).

The Canada Harbour marble deposit was investigated during the 1985 survey and chemical and physical analyses of chip samples collected from the deposit showed it to be one of two white marble prospects, located in the Canada Bay–Roddickton area, which warranted follow-up work. The other prospect, located at Penny's Pond near Roddickton, was selected for a follow-up program that included drilling. This led to the delineation of high-purity white marble deposits at Penny's Pond and Coles Pond (Howse, 1987), both of which have been awarded to industry for further assessment and development.

This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

The purpose of this paper is to comprehensively summarize the geology of the Canada Harbour deposit and compare its chemical and physical features to other white marble deposits in the region.

Location, Access, and Land Status

Canada Harbour, once the site of a now abandoned fishing settlement, is located at the entrance to Canada Bay, on the eastern side of the Great Northern Peninsula (Figure 1). There are no roads to the harbour. Englee, a major fish-processing centre for the region, is located 5 km north across the bay and is linked to the province's highway system. Canada Harbour is a sheltered inlet with deep water suitable for large ships (Plate 1). However, except for one or two small fishing wharves used by local fishermen, there are no docking facilities available in the harbour.

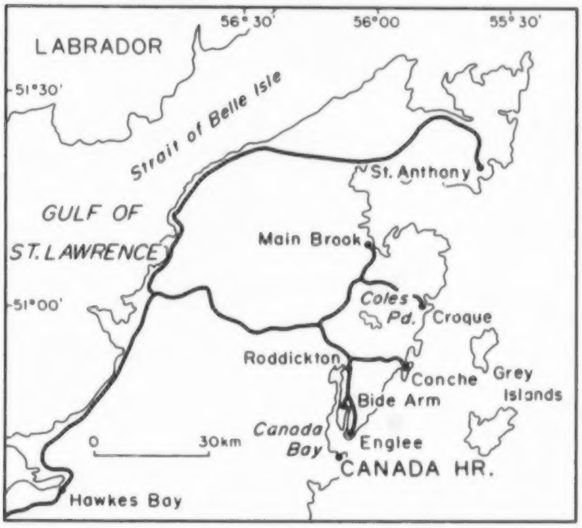

Figure 1. *Location of Canada Harbour, Newfoundland.*

The Canada Harbour marble deposit is encompassed by a Fee Simple Grant (Volume 1, Folios 134, 135), which was granted in 1919 to the Colonial Mineral and Trading Company, an English firm. The property is now held by Extender Minerals Limited of Mississauga, Ontario.

REGIONAL GEOLOGICAL SETTING

Deposits of high-purity white marble have been documented within a continuous belt extending from Canada Bay in the south to an area west of Croque, a distance of more than 50 km (Figure 2). The marble deposits occur in the Lower Paleozoic carbonate rocks, mainly within the Lower Ordovician St. George Group, where they have been deformed and metamorphosed in a series of thrust sheets beneath the Hare Bay Allochthon (Knight, 1987). In the Canada Harbour region, the platformal rocks are narrowly wedged between Precambrian crystalline basement on the west, and the structurally overlying allochthonous rocks to the east (Figure 3). The white marble zone closely parallels the western margin of the allochthon. The precise stratigraphic position of the Canada Harbour marble is uncertain. Knight (personal communication, 1988) stressed the need for further stratigraphic studies in that area of Canada Bay, and tentatively correlated the marbles with the Watts Bight Formation of the Lower Ordovician St. George Group.

Plate 1. *View of Canada Harbour looking northeast from the main quarry.*

DESCRIPTION OF THE DEPOSIT

The white marble deposit is exposed on Marble Ridge (Plate 2) in the main quarry, an open cut about 42 m by 24 m (Figure 4). A much smaller open cut (Quarry No. 2), which is located near the crest of the ridge, was examined but not sampled during the present study. In the main quarry, the white marble is gradationally overlain by blue-grey marble and dolomite. The deposit is structurally complex and bedding features are not apparent. Bain (1937) showed the white marble as occupying the core of an upright anticline, with the blue grey marble on its flanks. The thickness of the white marble unit is not known, but in the main quarry, a zone of approximately 35 m was measured and sampled. This measurement allows for the irregular layout (in plan) of the quarry face (Figure 4).

The marble is generally white, but is also ivory or cream and, in places has a distinctive pink hue. It is extremely fine grained and breaks with a conchoidal fracture. Some of the whiter stone has a porcelaineous texture and breaks much more irregularly. This feature seems to be more pronounced in the small or open-cut quarry.

Impurities in the marble include chlorite and sericite, which occur as thin coatings along joints and bedding planes.

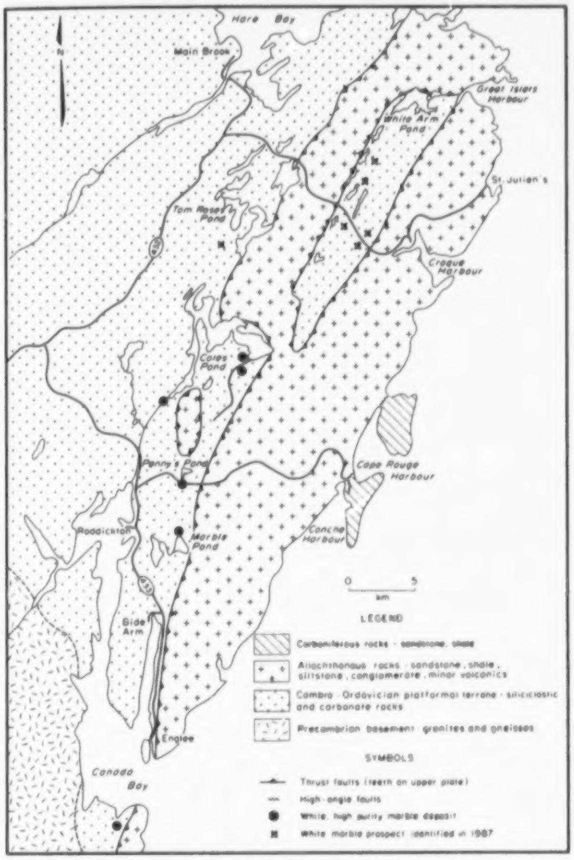
N
Hare Bay
Main Brook
Great Islets Harbour
White Arm Pond
St. Julien's
Tom Rose's Pond
Croque Harbour
Coles Pond
Penny's Pond
Cape Rouge Harbour
Conche Harbour
Roddickton
Marble Pond
Bide Arm
Englee
Canada Bay
0
5
km
LEGEND
Carboniferous rocks - sandstone, shale
Allochthonous rocks - sandstone, shale, siltstone, conglomerate, minor volcanics
Cambro - Ordovician platformal terrane - siliciclastic and carbonate rocks
Precambrian basement - granites and gneisses
SYMBOLS
Thrust faults (teeth on upper plate)
High-angle faults
White, high purity marble deposit
White marble prospect identified in 1987

Figure 2. *Geological elements of the Canada Bay–Hare Bay region showing location of white marble deposits. Geology simplified after Knight (1987) and Stouge (1983a,b).*

Quartz is usually present in only small amounts, however, where bands of dolomitic marble occur, the silica content is high. At least four such bands ranging up to 4 m in width are found in the face of the main quarry.

Barron (1967), identified two diabase dykes in exploration trenches that were dug across the white marble zone on Marble Ridge. The dykes have irregular widths ranging up to 4.5 m in width, are steeply dipping and cut the marble zone at 10 to 30°. They have porphyritic chill boundaries but the surrounding marble is relatively unaltered. The trenches are now refilled and thus inaccessible. However, this author noted a basic dyke on the shore of Canada Harbour about 300 m northeast of the main quarry, which is up to 4 m in width, and cuts the blue marble on the east flank of the white marble zone. Its 30° strike indicates it would intersect the white marble in the Marble Ridge area, if it were continuous.

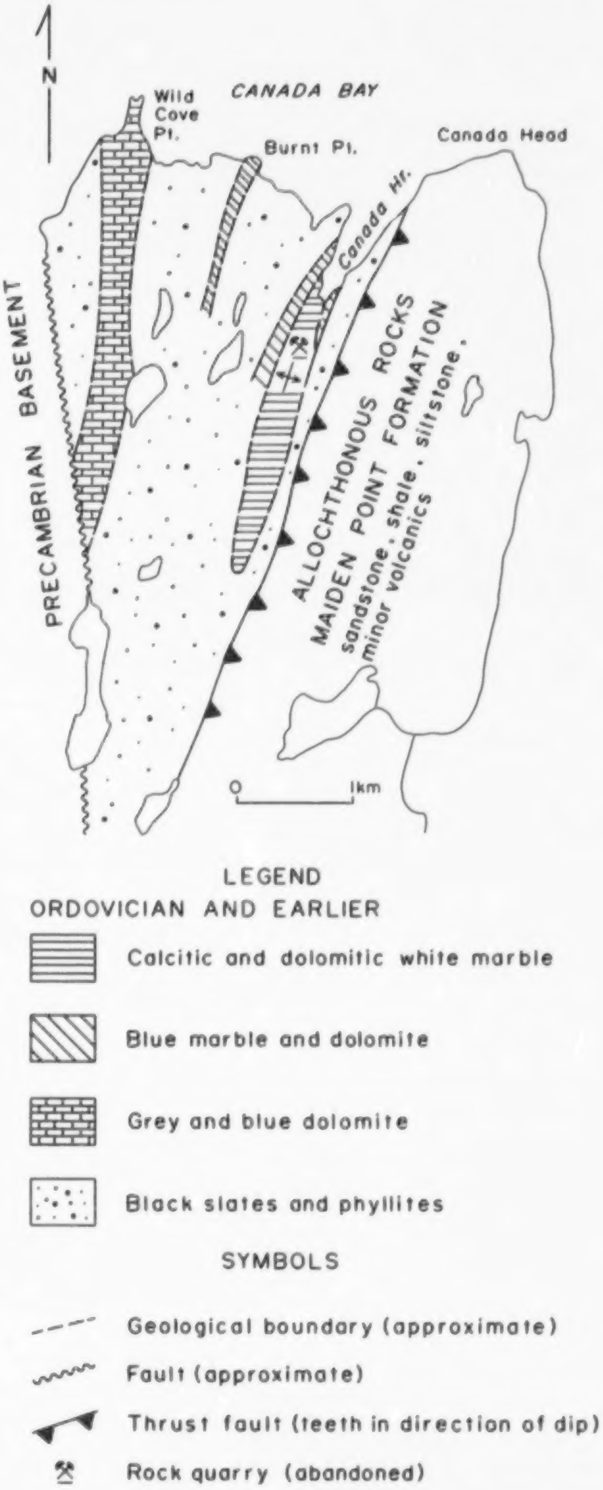
N
CANADA BAY
Wild Cove Pt.
Burnt Pt.
Canada Head
Canada Hr.
PRECAMBRIAN BASEMENT
ALLOCHTHONOUS ROCKS
MAIDEN POINT FORMATION
sandstone, shale, siltstone, minor volcanics
0
1km
LEGEND
ORDOVICIAN AND EARLIER
Calcitic and dolomitic white marble
Blue marble and dolomite
Grey and blue dolomite
Black slates and phyllites
SYMBOLS
Geological boundary (approximate)
Fault (approximate)
Thrust fault (teeth in direction of dip)
Rock quarry (abandoned)

Figure 3. *Geological sketch map of the Canada Harbour area; modified after DeGrace, 1974.*

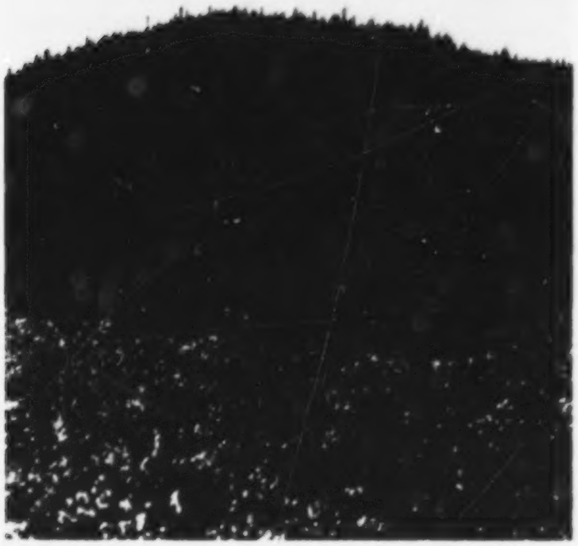

Plate 2. *Marble ridge as viewed from the abandoned settlement of Canada Harbour.*

Reserve Potential

Given the present information available on the Canada Harbour marble it is difficult to estimate the size of the deposit. The marble is exposed on the coast at Canada Harbour, in the open cuts on Marble Ridge and in a few scattered outcrops south of the ridge, for a total length of approximately 2.5 km. The area is covered with a thin, but very pervasive blanket of till, therefore, it is not presently known if the zone is continuous along its strike. For practical purposes, only Marble Ridge and the area south can be considered as having reserve potential. Therefore, if one assumes the white marble zone is continuous south of the main quarry for 2 km, and further assumes a width of 35 m, and a quarryable depth of 30 m, a tonnage of about 4.6 million tonnes is indicated using a density (specific gravity) of 2.17 g/cm^3. The actual mining depth that might be realized by a potential developer is a major uncertainty. The presence of diabase dykes of unknown dimensions in the deposit is a major concern that underscores the need for further work.

It is clear that a systematic drilling program is essential in order to confirm and accurately measure the tonnages and grades of the deposit.

Chemistry and Dry Brightness of the Marble

The chemical analyses show that approximately 20 percent of the white marble unit in the main quarry is dolomite (Table 1). The dolomite beds are high in impurities, especially quartz, and are intimately interbanded with the calcitic marble. The calcitic marble is fairly pure with the main contaminant, silica, averaging 1.03 percent, and alumina and iron totalling well below one half percent. The average grade, making no distinction between the dolomite and calcite, is about 89 percent $CaCO_3$, 2.5 percent silica, and $MgCo_3$ at just over 6.2 percent. Alumina and iron together, total less than one half percent.

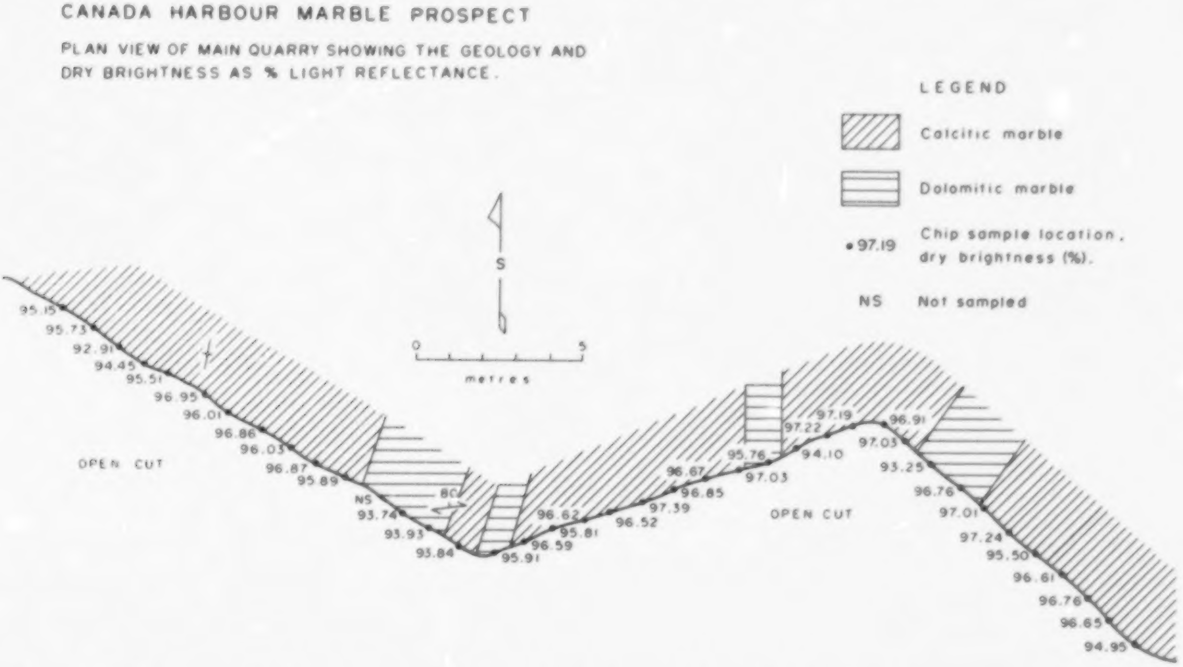
CANADA HARBOUR MARBLE PROSPECT
PLAN VIEW OF MAIN QUARRY SHOWING THE GEOLOGY AND DRY BRIGHTNESS AS % LIGHT REFLECTANCE.
LEGEND
Calcitic marble
Dolomitic marble
• 97.19 Chip sample location, dry brightness (%).
NS Not sampled
S
0
5
metres
OPEN CUT
95.15
95.73
92.91
94.45
95.51
96.95
96.01
96.86
96.03
96.87
95.89
NS
93.74
93.93
93.84
80
95.91
96.59
95.81
96.62
96.52
97.39
96.85
96.67
97.03
95.76
94.10
97.22
97.19
96.91
97.03
93.25
96.76
97.01
97.24
95.50
96.61
96.76
96.65
94.95
OPEN CUT

Figure 4. *The plan view of the main quarry, showing the geology and the dry brightness as a percentage of light reflectance.*

Table 1: Chemical analyses and dry brightness tests for chip samples from the Canada Harbour white marble deposit. Chips represent one-metre intervals on the main quarry face going from east to west (see Figure 4).

	PERCENT						
Sample No.	SiO_2	Al_2O_3	Fe_2O_3	$MgCO_3$	$CaCO_3$	DB*	ROCK TYPE
5945898	0.62	0.14	0.07	0.54	96.12	95.15	calcitic marble
5945899	0.68	0.15	0.08	0.48	97.19	95.73	calcitic marble
5945901	4.67	0.83	0.22	1.00	91.08	92.91	calcitic marble
5945902	0.67	0.07	0.05	0.54	98.08	94.45	calcitic marble
5945903	0.60	0.12	0.05	0.50	98.61	95.51	calcitic marble
5945904	3.28	0.54	0.12	0.65	93.70	96.95	calcitic marble
5945905	1.31	0.41	0.14	0.71	97.35	96.01	calcitic marble
5945906	0.39	0.06	0.06	0.46	98.15	96.86	calcitic marble
5945907	0.82	0.19	0.09	0.69	96.69	96.03	calcitic marble
5945908	0.72	0.10	0.05	0.54	98.18	96.87	calcitic marble
5945909	5.49	0.48	0.25	32.52	60.45	95.89	dolomitic marble
5945911	31.35	4.84	1.40	24.07	35.12	**	dolomitic marble
5945912	2.99	0.41	0.27	39.96	56.67	93.74	dolomitic marble
5945913	4.38	0.34	0.28	40.56	54.22	93.93	dolomitic marble
5945914	0.73	0.16	0.06	0.79	97.13	93.84	calcitic marble
5945915	4.39	0.67	0.29	23.68	70.17	95.91	dolomitic marble
5945916	0.52	0.07	0.03	1.10	96.80	96.59	calcitic marble
5945917	0.38	0.05	0.03	0.52	97.37	95.81	calcitic marble
5945918	0.32	0.04	0.03	0.54	97.03	96.62	calcitic marble
5945919	0.50	0.10	0.05	4.26	93.29	96.52	calcitic marble
5945921	0.38	0.06	0.04	0.48	98.13	97.39	calcitic marble
5945922	0.53	0.10	0.04	0.73	97.33	96.85	calcitic marble
5945923	1.88	0.31	0.14	4.14	91.69	96.67	calcitic marble
5945924	0.40	0.07	0.04	0.52	97.10	97.03	calcitic marble
5945925	7.81	1.41	0.64	23.78	63.85	95.76	dolomitic marble
5945926	2.99	0.89	0.30	8.57	86.77	94.10	calcitic marble
5945927	0.64	0.11	0.06	0.73	97.53	97.22	calcitic marble
5945928	0.48	0.07	0.05	0.58	99.48	97.19	calcitic marble
5945929	0.67	0.14	0.09	0.81	97.44	96.91	calcitic marble
5945931	0.64	0.12	0.07	0.77	97.77	97.03	calcitic marble
5945932	5.14	0.17	0.16	10.74	72.46	93.25	dolomitic marble
5945933	4.03	1.00	0.38	13.17	79.16	96.76	dolomitic marble
5945934	0.49	0.12	0.06	0.48	97.49	97.01	calcitic marble
5945935	0.45	0.08	0.04	0.42	96.56	97.24	calcitic marble
5945936	1.53	0.24	0.12	0.63	95.55	95.50	calcitic marble
5945937	0.91	0.32	0.10	0.61	96.07	96.61	calcitic marble
5945938	0.78	0.25	0.10	0.65	96.05	96.76	calcitic marble
5945939	2.14	0.45	0.19	0.75	94.43	96.65	calcitic marble
5945941	0.83	0.24	0.11	1.02	96.23	94.95	calcitic marble
AVERAGE	2.50	0.40	0.16	6.24	89.08	95.96	combined average
AVERAGE	1.03	0.25	0.10	1.13	96.21	96.16	calcitic marble
AVERAGE	8.19	1.16	0.45	26.06	61.51	94.75	dolomitic marble

NOTE: * Dry brightness (percent light reflectance) using the tri-stimulus method as described in A.S.T.M. test method E-97); ** not analyzed; Tests for brightness by I.M.D. Laboratories Limited (1986).

The dry brightness of the marble, a critical factor in determining the marble's suitability as an industrial filler, averages 95.96 percent. Excluding the dolomite beds, which average a surprisingly high 94.75 percent, the samples show an average brightness of 96.16 percent. Seven of the samples (18 percent of the total) had values in excess of 97 percent, all within the calcitic marble. This shows that although there is a significant brightness difference between the dolomite and calcite, dilution in brightness due to the dolomite does not seriously affect the overall brightness of the marble.

Table 2. Average chemical analyses and dry brightness from the Coles Pond (North) and Penny's Pond deposits.

Coles Pond (North) Drill Core Analyses

DDH	Metreage	SiO_2+	Al_2O_3+	Fe_2O_3+	$MgCO_3$+	$CaCo_3$+	DB*	T
14	1.52-27.45	1.70	0.44	0.15	0.65	95.94	96.00	276,000
15	2.14-38.13	1.70	0.45	0.14	0.67	96.36	95.63	229,500
16	2.14-38.13	1.46	0.34	0.12	0.63	96.72	96.14	248,800
17	2.44-32.02	1.57	0.35	0.11	0.63	96.26	96.38	183,000
18	3.66-35.68	1.46	0.47	0.12	0.63	97.01	95.88	171,000
19	2.74-34.46	1.06	0.23	0.11	0.56	97.93	95.61	138,862
	overall average	1.52	0.39	0.13	0.63	96.60	95.95	

Penny's Pond Drill Core Analyses

DDH#	Metreage	SiO_2+	Al_2O_3+	$Fe_2O_3$12+	$MgCO_3$+	$CaCo_3$+	DB*	T
1	2.14-51.85	1.42	0.38	0.14	1.08	93.99	95.66	260,000
2	3.05-51.85	1.62	0.39	0.13	0.65	96.36	96.01	308,100
3	32.02-57.95	1.88	0.36	0.13	0.98	96.34	96.02	347,700
4	3.05-59.48	1.49	0.35	0.15	0.65	96.61	95.82	386,250
5	1.52-69.54	1.48	0.35	0.12	0.86	97.02	96.08	483,200
11	2.75-10.68	1.09	0.27	0.09	0.56	96.95	96.00	14,400
	overall average	1.58	0.36	0.13	0.84	96.58	95.83	

* Dry Brightness Index
+ Weight Percent
T Metric Tonnes

Tests for brightness by I.M.D. Laboratories (1987).

COMPARISONS BETWEEN THE CANADA HARBOUR AND RODDICKTON–CROQUE DEPOSITS

There are obvious similarities between the Canada Harbour marble and that found in the Roddickton area, i.e., Penny's Pond and Coles Pond. In each case, the marble is very fine grained and slightly cleaved in weathered outcrops, a feature not apparent in drill core (Penny's Pond and Coles Pond) or in quarry faces (Canada Harbour). Chlorite is ubiquitous as a minor contaminant occurring as thin films along fractures and in bands of carbonate schist several centimetres thick. Other impurities include very minor amounts of sericite along some fractures, and rare quartz. Detailed sampling, however, shows that there are significant differences in the chemistry of the deposits. Chemical analysis of the drill log for both the Penny's Pond and Coles Pond deposits (Howse and Delaney, 1987) show that the white marble is almost 100 percent calcitic (Table 2). The analyses of continuous drill core (at 1.52-m intervals), show a complete absence of dolomite, and the average $MgCO_3$ content for the two deposits is 0.74 percent. By contrast, about one fifth of the white marble in the Canada Harbour main quarry is dolomite and the average $MgCO_3$ content is over 6 percent. The silica content for the Canada Harbour marble (2.5 percent), is about 1 percent higher than the average at Penny's Pond and Coles Pond. There is no significant difference in the alumina and iron content, both of which total less than 0.5 percent in the three deposits.

There is little variation in the average marble brightness for the three deposits. For the Canada Harbour deposit, the 95.96 percent reflectance is very close to the 95.83 and 95.95 percentages, respectively, for the Penny's Pond and Coles Pond deposits. Locally, some bands of Canada Harbour marble, however, have a superior brightness (>97 percent light reflectance). The brightness of the dolomitic marble is surprisingly high, ranging between 93.74 and 95.91 percent (average 94.74). The brightness dilution from the interbanded dolomite does not appear to significantly lower the average for the deposit.

There is evidence to suggest that the Canada Harbour marble may closely resemble the white marble in the White Arm Window (Croque area). For example, the Powerline prospect (Howse, 1988) on the eastern side of the White Arm Window consists of intimately interbedded calcitic and dolomitic marble of high brightness (Table 3). The brightness of the dolomitic marble at 93.03 percent and the calcitic marble at 95.43 compares favourably to the Canada Harbour stone. The marble deposit of Aurion Minerals (W. MacQuarrie, personal communication, 1988), located just

Table 3: Chemical analyses and dry brightness tests for representative chip samples from the Powerline marble occurrence near Croque.

	PERCENT						
Sample No.	SiO_2	Al_2O_3	Fe_2O_3	$MgCO_3$	$CaCO_3$	DB*	ROCK TYPE
5948119	1.84	0.29	0.07	0.75	96.96	94.33	calcitic marble
5948121	0.42	0.15	0.06	0.48	98.68	96.54	calcitic marble
5948122	11.03	0.98	0.49	33.69	52.26	93.03	dolomitic marble
average	4.43	0.47	0.20	11.64	82.63	94.63	

NOTE: * Dry brightness; Tests for brightness by I.M.D. Laboratories (1988)

north of the Powerline prospect, has similar characteristics including bands of superior dry brightness calcitic marble interbanded with dolomitic sections.

CONCLUSION

The Canada Harbour white marble prospect has several features that make it economically attractive. It is located at tidewater, close to shipping lanes, and although relatively isolated it can be easily reached by boat across Canada Bay from Englee, a prosperous fishing centre. The deposit is located on a ridge having natural quarrying advantage, and there is ample room for development.

The present assessment indicates the Canada Harbour marble is less pure, chemically, than the Roddickton deposits. Nevertheless, the dry brightness of the deposit is excellent. This is a critical factor in determining its suitability for industrial uses that require premium quality filler. It seems certain, however, that for such applications the marble would require secondary processing in order to reduce impurities to an acceptable level.

The Canada Harbour deposit needs to be accurately delineated by diamond drilling. Such a program should be designed to trace the marble southward from the area of the main open cut on Marble Ridge. The present evidence suggests that this area may be underlain by a significant reserve of high-quality stone, but this can only be confirmed by further work.

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LABRADORITE OCCURRENCES NORTH OF DAVIS INLET

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Mineral Deposits Section

ABSTRACT

Newly reported labradorite occurrences in the Tunungayualok Island area, north of Davis Inlet, have been evaluated as a source of both gem-quality stone and dimension stone. The labradorite occurs in breccia zones and dykes within a large body of anorthositic rocks, in which, the crystals are typically fractured, making it difficult to extract from the breccia. In several prospects, the percentage of visibly chatoyant labradorite is up to 15 percent.

No potential dimension-stone sites were observed within the study area.

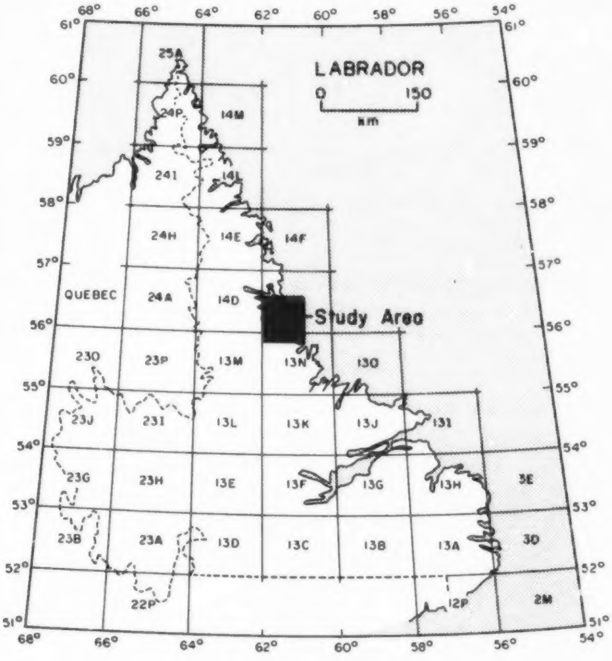
INTRODUCTION

Labradorite was discovered in the Nain area (Figure 1) in the 1700's and is presently quarried at Tabor Island, about 15 km south of Nain. It is used locally in the making of jewellery and other handicrafts and a small amount is sold to lapidary shops outside of Labrador. A survey of labradorite occurrences was carried out by the Newfoundland Department of Mines in 1979 (Watson, 1980), and 7 major prospects and 20 or more minor prospects were identified, in the coastal area between Nain and Kamarsuk (NTS 14C/5,6,11,12) (Figure 2).

A new labradorite prospect was reported to the Newfoundland Department of Mines in the spring of 1988, by Bert Saunders of Davis Inlet. The prospect is located on Tunungayualok Island, 16 km north of Davis Inlet (Figure 2) and was subsequently visited by one of the authors (H. Montague) and based upon the initial evaluation, an additional two-week survey of the area was carried out in August 1988, to investigate the economic potential for additional labradorite prospects.

GEOLOGY

The surveyed area consists of approximately 300 km² of anorthositic rocks, which have been mapped by Wiebe (1974, 1975) (Figure 3). Archean basement is exposed on the eastern side of Tunungayualok Island, and on islands to the north and south. Younger adamellite (mostly quartz monzonite and granodiorite) form the western margin of the anorthositic rocks, and adamellite and small dioritic bodies and dykes, occur within the anorthosite rocks.



INDEX MAP

Figure 1. *Location of study area.*

The anorthositic rocks are divided into three main units by Wiebe (1974, 1975), although all the three units are internally heterogeneous and extremely variable in texture. The three units as determined by Wiebe *(op. cit)* are: 1) anorthosite—which is generally dark grey, and ranging from massive anorthosite and relatively even-textured 2-cm plagioclase crystals, to seriate anorthosite and plagioclase crystals 2 to 20 cm in length; 2) leuconorite—which consists of anorthositic rocks having a higher colour index (mostly 15 to 25), due to the increased amount of orthopyroxene; and 3) megabreccia—a transition zone between units 1 and 2 with

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This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

[1] Mineral Resource Management Branch

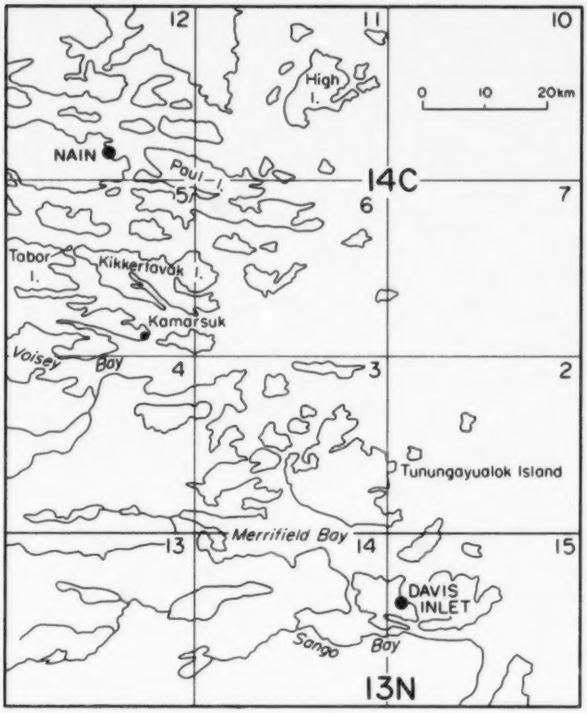

Figure 2. *Location map for area between Davis Inlet and Nain.*

angular to rounded blocks of anorthosite (several centimeters to tens of metres in size) within a leuconoritic matrix.

LABRADORITE

Labradorite prospects on Tunungayualok Island differ from those in the Tabor Island–Kikkertavak Island area in the north. In the latter areas, they consist of extremely coarse grained pegmatitic pods of tightly packed labradorite crystals and minor pyroxene. The labradorite prospects in the Tunungayualok Island area consist of breccia zones or dykes, containing angular to rounded labradorite crystals (Plate 1). The crystals range in size from 1 to 35 cm and are typically fractured, often cut by thin white feldspar veins. The degree of chatoyancy visible in outcrops, varies considerably, and in some prospects ranges up to 15 percent; the typical colours are deep-blue, blue-green to yellow and frequent colour zoning. The percentage of matrix in the breccia zones is very varied as well, ranging from white to grey. The white matrix, which weathers a distinctive yellow to orange-brown, consists of crushed labradorite crystals, 1 to 10 mm in diameter, often having accessory pyroxene, magnetite, pyrite, and less commonly quartz and pale-orange feldspar. In some prospects, the matrix content is as high as 20 percent, but in others it is less than 5 percent. The breccia zones are up to several hundred metres wide in shoreline exposures, and are generally located within the megabreccia unit. The leuconorite and norite units also contain several breccia zones and numerous dykes ranging from 1- to 5-m in width. The following is a descriptive account of some of the better labradorite prospects in the Tunungayualok Island area.

Kayutak Bay

Kayutak Bay is located on the northwestern side of Tunungayualok Island (Figure 3). At the east end of Kayutak Bay, there is a moderately steep talus slope that stretches approximately 75 m from the outcrop down to the ocean. The talus slope contains hundreds of blocks of rock ranging in size from 10 cm to 2 m. The blocks are almost all composed of labradorite breccia. Approximately 75 percent of the breccia has a white matrix, and the remainder has a dark grey matrix. Many of the blocks are partially iron-stained. The breccia contains rounded to angular chatoyant labradorite crystals ranging in size from 1 to 30 cm, (average size approximately 4 cm), and has a matrix content of 5 to 15 percent. The crystals are generally fractured, but there are many of high quality that have excellent colour, colour zoning, and twinning. Some of the colour-zoned crystals show repeated zoning in an octagonal pattern. There are an abundance of labradorite fragments in the intertidal zone at the base of the talus slope. The majority of these, however, are fractured, and would not be suitable for high-quality jewellery. Despite the availability of loose blocks, the dimension-stone potential is limited, as most blocks are less than 1 m in diameter, fractured, iron-stained, and irregularly shaped.

Nuvudluktok Bay

Nuvudluktok Bay, on the west side of Tunungayualok Island, is a large inland bay having many smaller coves and several islands and is well protected from the sea with a very narrow entrance. The easternmost of the two large islands in the bay is underlain by labradorite breccia and megabreccia. At the west end of this island (Figure 3), there are large exposures of breccia having a high matrix content and large chatoyant labradorite crystals, commonly up to 15 cm in length. The crystals display a variety of colours and colour zoning, but they are typically fractured and would be difficult to extract for jewellery. There are many smaller occurrences of labradorite breccia exposed along the shoreline of Nuvudluktok Bay, and several outcrops of medium to coarse grained anorthosite exhibiting 5 to 10 percent chatoyant labradorite. However, the anorthosite outcrops are too small and fractured to be considered as dimension-stone prospects.

South Tunungayukuluk Island

This prospect, on the flat-lying, wave-washed northeast corner of the island (Figure 3), is a good example of Wiebe's (1974) megabreccia unit. Rounded inclusions of medium grained anorthosite, up to a metre in diameter, are well exposed in an orange weathering, white matrix that also contains numerous chatoyant labradorite crystals (Plate 2). The labradorite crystals are up to 45 cm in diameter, but generally average about 5 cm. The larger crystals are rounded and display a beautiful variety of colours and colour zoning,

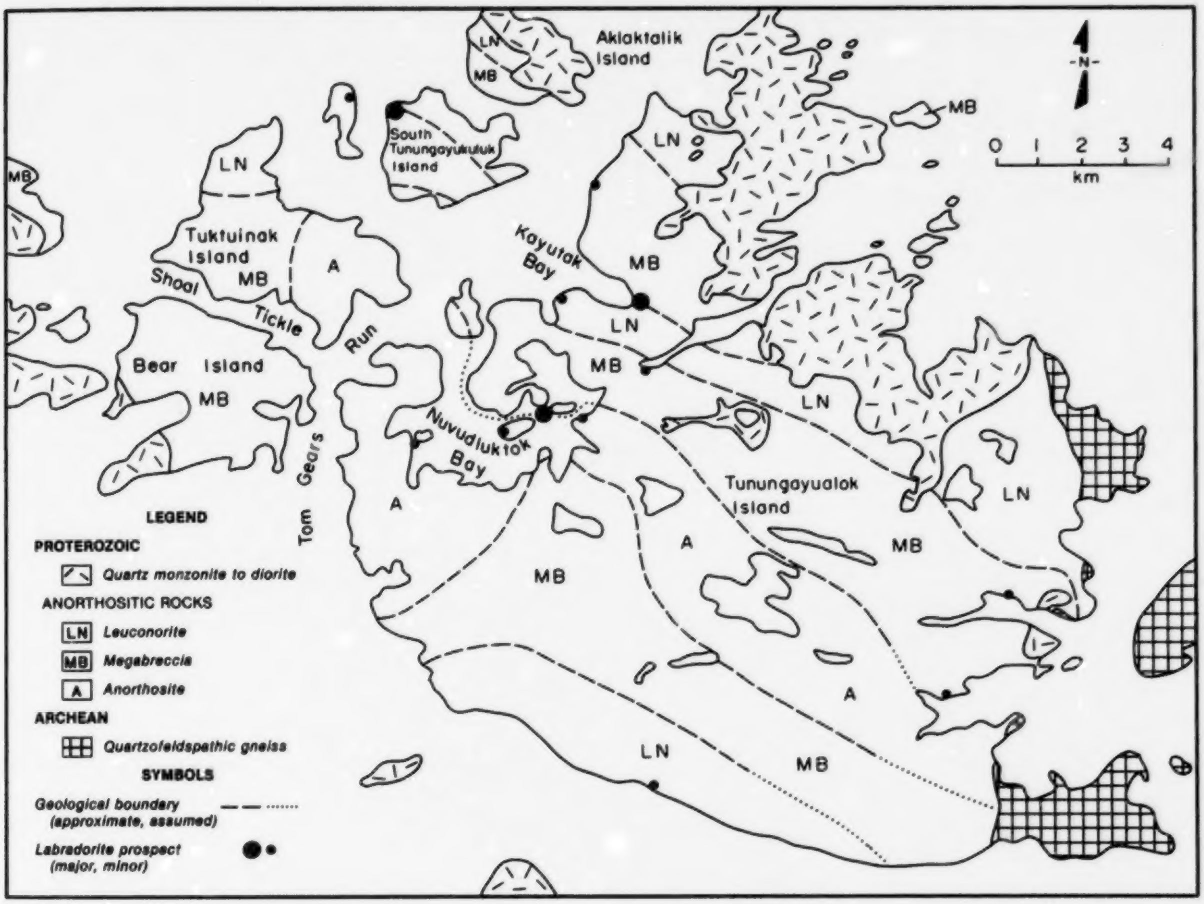
Akiaktalik Island
South Tunungayukuluk Island
Tuktuinak Island
Shoal Tickle
Run
Bear Island
Tom Gears
Kayutak Bay
Nuvudluktok Bay
Tunungayualok Island
LN
MB
A
-N-
0 1 2 3 4
km
LEGEND
PROTEROZOIC
Quartz monzonite to diorite
ANORTHOSITIC ROCKS
LN Leuconorite
MB Megabreccia
A Anorthosite
ARCHEAN
Quartzofeldspathic gneiss
SYMBOLS
Geological boundary (approximate, assumed)
Labradorite prospect (major, minor)

Figure 3. *Geology of Tunungayalok Island (after Wiebe, 1975).*

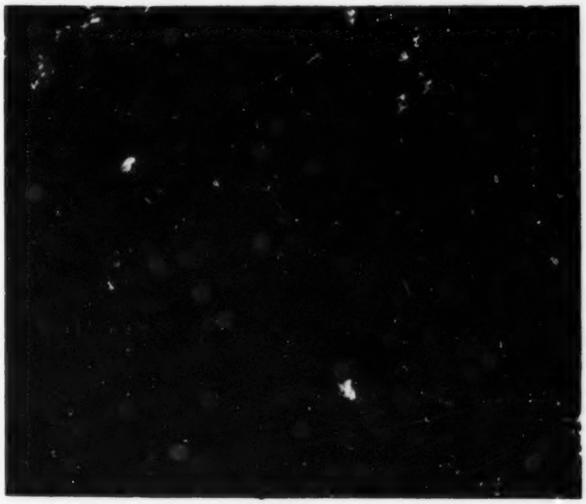

Plate 1. *Fractured outcrop of labradorite breccia containing crystals of labradorite ranging from 2 to 20 cm in length.*

Plate 2. *Coastal outcrop from South Tunungayukuluk Island prospect showing large rounded anorthosite inclusions and 2- to 10-cm labradorite crystals in orange-brown weathered matrix. (Note pen for scale.)*

but are extensively fractured and cut by thin white feldspar veins, which cut across both the matrix and inclusions.

CONCLUSIONS

Labradorite occurrences in the Tunungayualok Island area have a low potential for gem-quality stone, when compared to existing labradorite quarries in the Nain area. In contrast to the very coarse grained pegmatitic labradorite occurrences in the Nain area, these recently investigated occurrences consist of medium- to coarse-grained labradorite crystals, which occur within breccia zones. The labradorite crystals are generally fractured, and may contain impurities or be cut by thin white veins. The matrix content of the breccia zones is 5 to 15 percent. The individual crystals, which are generally much smaller than those found in the pegmatitic occurrences, would have to be cut out individually to obtain gem-quality stone, providing good sites with abundant unfractured crystals could be located. Alternatively, loose pieces of labradorite could be scavenged in the vicinity of the larger breccia zones, but these weathered pieces are often unsuitable.

The dimension-stone potential of the labradorite breccia is limited by the generally small block sizes that could be quarried. Most breccia zones are moderately to strongly fractured and seldom would it be possible to quarry blocks that are larger than 1 m^3. Smaller blocks of labradorite would be marketable if a high enough percentage of chatoyant labradorite were visible on a cut surface. Preliminary results of slabbed samples indicate that this would be rare. Whereas the irregular surface of several outcrops visible in the field were quite spectacular when washed with water, a smooth, cut surface never displays as high a percentage of chatoyant labradorite. This is analagous to previous studies carried out by the Newfoundland Department of Mines using labradorite aggregate in concrete work (Meyer and Dean, 1987). It was found that the smooth surface of a terrazzo tile never displayed as high a percentage of chatoyant labradorite crystals as the rough, irregular surfaces of a precast concrete panel, which preserves the natural cleavage faces of the labradorite crystals.

ACKNOWLEDGMENTS

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RARE-METAL TARGETS IN INSULAR NEWFOUNDLAND

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ABSTRACT

The extensive infrastructure, less harsh weather conditions and lower exploration costs make insular Newfoundland a more economically attractive place than Labrador to find and develop rare-metal deposits. Yet, to date, the only large rare-metal deposits discovered in this province, the Strange Lake deposit and Mann # 1 showing, occur in Labrador. Little, if any, rare-metal exploration has been carried out in insular Newfoundland.

A model of rare-metal mineralization in high-level felsic peralkaline rocks has been developed from the study of the Strange Lake deposit, the Mann # 1 showing and several other showings in Labrador. The model outlines four major geological settings for rare-metal mineralization in high-level felsic peralkaline rocks. These analogous geological settings can be sought in insular Newfoundland, to assist in rare-metal exploration programs.

Application of the model to insular Newfoundland indicates that eight complexes or regions, contain favourable host rocks for rare-metal mineralization; i.e., felsic extrusive and/or intrusive rocks that are peralkaline in nature or affinity. These exploration targets are: Louil Hills complex–Bull Arm Formation, Cross Hills complex–Mooring Cove Formation, Topsails complex–Sheffield group, St Lawrence Granite and related rocks, Kings Point complex, Grand Beach complex, Seal Island Bight granite, and the Hare Hills peralkaline rocks.

INTRODUCTION

Recent geological and geochemical studies of the rare-metal mineralization within the Strange Lake peralkaline granite, the Red Wine intrusive suite and the Letitia Lake Group (Miller, 1986, 1987) of Labrador (Figure 1), has led to the development of an empirical metallogenic model for these felsic peralkaline rock-hosted deposits. The model can be successfully applied to other similar rare-metal deposits (e.g., Thor Lake, N.W.T., Trueman *et al.*, 1985; Jabal Sa'id, Saudi Arabia, Drysdale *et al.*, 1984; Hackett, 1986), and thus has application to the exploration for undiscovered rare-metal mineralization. In this study, the model is used to constrain and define potential rare-metal targets on the Island of Newfoundland.

To date, the two significant rare-metal deposits discovered in Labrador are a result of exploration programs for uranium in the Strange Lake region (Miller, 1986), and for copper and zinc in the Mann #1 showing (Brummer, 1960). The fact that these exploration programs were not aimed at rare-metal mineralization is due to: 1) the very small number of companies focussing on rare-metal exploration versus the traditional commodities such as Fe, U, Au, Cu or Zn, 2) the low demand and small markets, and 3) the considerably higher cost of exploration programs in Labrador. Thus, the rare-metal discoveries in Labrador were 'fortuitous', rather than 'by design'. The development of an empirical model should help lower the costs, by assisting companies to focus on smaller areas for evaluation, which will better identify the potential that will attract rare-metal exploration companies to Province of Newfoundland and, in particular, insular Newfoundland.

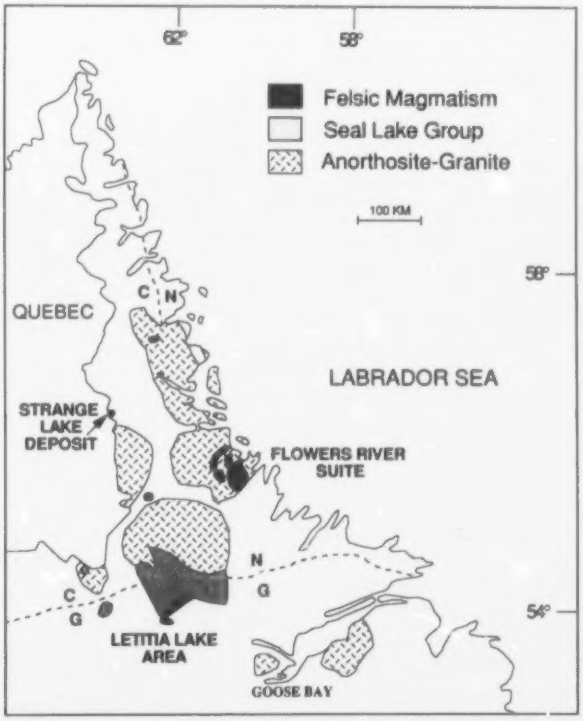

Figure 1. *Rare-metal showings in Labrador. Strange Lake Zr–Y–Be–Nb–REE deposit; Letitia Lake area: Mann #1, Mann #2 and Two Tom Lake showings; Flowers River suite–minor showings.*

This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

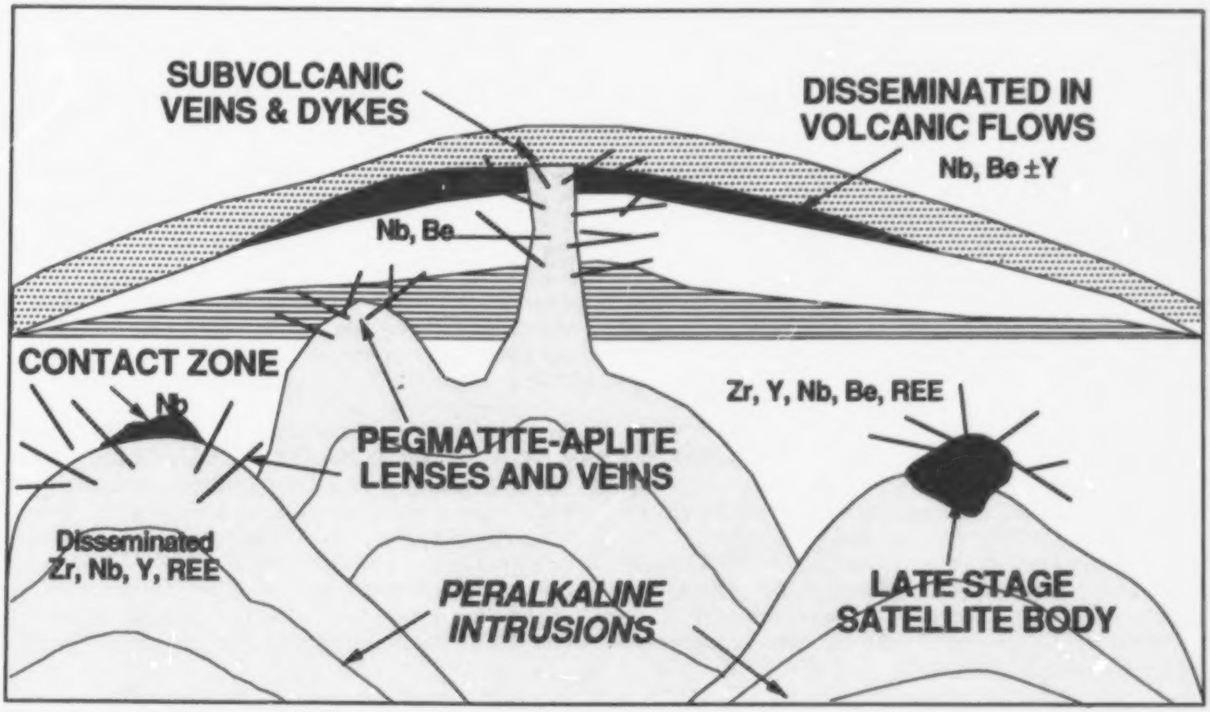
SUBVOLCANIC VEINS & DYKES
DISSEMINATED IN VOLCANIC FLOWS
Nb, Be ±Y
Nb, Be
CONTACT ZONE
Nb
Zr, Y, Nb, Be, REE
PEGMATITE-APLITE LENSES AND VEINS
Disseminated Zr, Nb, Y, REE
PERALKALINE INTRUSIONS
LATE STAGE SATELLITE BODY

Figure 2. *Rare-metal mineralization model: Setting 1—late-stage satellite bodies and associated pegmatite—aplite veins associated with unvented plutons (granite); settings 2 and 3—veins and disseminated mineralization in near vent volcanic rocks and subvolcanic equivalents (granite or syenite); setting 4—disseminated mineralization and contact zones in unvented plutons (granite and undersaturated syenites).*

When compared to Labrador, insular Newfoundland provides a more favourable environment for exploration and development. Exploration costs are much lower due to the extensively developed infrastructure, the larger workforce and the larger number of services available. These factors would also mean that unlike Labrador, significant discoveries of rare-metal mineralization are more likely to be developed. Thus, insular Newfoundland is a better place to find rare-metal deposits and to bring them into production.

GEOLOGICAL SETTINGS OF RARE-METAL MINERALIZATION

The metallogenic model developed for high-level peralkaline rock-hosted rare-metal mineralization is based on the metallogenic studies of the Strange Lake Zr—Y—Nb—Be—REE deposit (Miller, 1986), the Mann-Type Nb—Be±Y showings in the Letitia Lake area (Miller, 1987; Batterson and Miller, 1987), the mineralization observed in the undersaturated phases of the Red Wine instrusive suite (Miller, 1988) and the mineralization described in the Flowers River Igneous Suite (Hill, 1981; Miller, 1988). The general geological settings and the geochemistry of the mineralization in these Labrador-based examples are certainly not representative of all of the many different kinds of rare-metal deposits found world-wide, nevertheless, a search of the literature indicates that the major settings have been described in the model, and these are illustrated in Figure 2.

This model indicates that rare-metal mineralization, in general (i.e., complex mineralization including one or more of Zr, Y, REE, Nb, Ta, Be), occurs in anorogenic high-level peralkaline felsic intrusions and their near-vent extrusive equivalents (Miller, 1987). The following geological settings have been identified in Labrador:

1) as pegmatite—aplite dykes and late-stage roof zone phases in high-level peralkaline granite plutons that did not vent to the surface (e.g., Strange Lake peralkaline granite),

2) as pegmatite and pegmatite—aplite dykes in the roof zone of vented peralkaline granites and as disseminated mineralization in the near-vent extrusive equivalents, (e.g., Flowers River intrusive suite, Shallow Lake peralkaline granite),

3) as subvolcanic veins associated with peralkaline quartz syenites and as disseminated mineralization in the associated peralkaline trachytes within the near-vent environment (e.g., Mann #1 and Two Tom Lake deposits),

4) as disseminated mineralization within high-level undersaturated peralkaline complexes (e.g., Red Wine intrusive complex of Curtis and Currie, 1981).

Table 1. Minimum and maximum values of rare-metal concentrations in rare-metal host rocks in Labrador.

Setting	Host	Zr	Y	Nb	Be	U
Setting 1	Strange Lake granite	2000	600	250	30	7
		5000	1500	500	80	30
Setting 2	Flowers River granite	600	100	50	-	2
		1100	150	150	-	6
Setting 2	Flowers River volcanics	1000	100	40	-	2
		8000	600	300	-	6
Setting 3	Letitia Lake syenites	700	50	50	7	1
		3600	300	400	50	12
Setting 3	Letitia Lake Group	200	60	20	5	3
		2000	200	800	100	6
Setting 4	Red Wine nepheline syenite	500	50	90	-	1
		5000	200	300	-	14

Sources of data: Strange Lake granite—Miller, 1986; Flower River volcanics—Hill, 1981; Letitia Lake Group—Miller, 1987; Flowers River granite—Hill, 1981; Letitia Lake syenites—Miller, 1987; Red Wine nepheline syenite—Curtis and Currie, 1981; rare-metal values in p.p.m.

Setting numbers 1 and 3 contain significant tonnages of high-grade mineralization (Miller, 1988). High values of rare metals have been found in settings 2 and 4, although significant tonnages have not been outlined.

These same settings have also been shown to contain significant rare-metal mineralization in other parts of the world such as the Northwest Territories (Thor Lake, Trueman *et al.*, 1985), Saudi Arabia (Drysdale *et al.*, 1984; Jackson, 1986), Greenland (Industrial Minerals, 1988b,c) and Australia (Industrial Minerals, 1988a).

The Thor Lake deposit exhibits many of the characteristics of setting 1, although there are many complexities with regard to the genesis of this Be–Nb–Ta–Y–REE–Ga deposit (Trueman *et al.*, 1985; Birkett, 1988). The mineralization occurs in several zones, some associated with metasomatized granite pegmatite and some associated with highly altered peralkaline granite (Grace Lake granite) and amphibole ± olivine ± pyroxene syenite (Thor Lake syenite). There is no evidence to suggest that either the syenite or granite was vented to the surface, although the high-level nature of these rocks is clearly evident. The high degree of metasomatism in the Thor Lake deposit suggests a late-stage development of rare-metal mineralization, in which the late stage is not manifested dominantly by pegmatite–aplite or by vein formation, but by intense hydrothermal–metasomatic activity, possibly associated with pegmatite formation.

Several rare-metal deposits and showings associated with high-level peralkaline felsic rocks have been described from Saudi Arabia (Drysdale *et al.*, 1984). All of these examples are associated with granites or syenites, which are not vented to the surface (Jackson, 1986). One particularly well-studied example is the Jabal Sa'id pegmatite–aplite hosted Nb–Zr–Y–REE–Sn–Ta deposit (Hackett, 1986). The mineralization occurs within a composite pegmatite–aplite body, which is at the contact of a peralkaline granite. This mineralization corresponds to setting 1, and is very similar to the Strange Lake deposit.

Two separate exploration programs in Greenland (Industrial Minerals, 1988b, 1988c) are evaluating eudialyte-bearing peralkaline undersaturated phases of the Ilimaussaq felsic complex. Both properties are being evaluated for Zr–Y–REE deposits similar to the mineralization observed in the Red Wine undersatured rocks and setting 4.

There is very little published geological information on the Brockman Nb–Ta–Y–Zr–REE deposit in Australia (Industrial Minerals, 1988a). Yet, it appears that the mineralization is disseminated in trachytic flows and tuffs of the Brockman volcanics. Mineralogical and other studies of the volcanic host rocks (A. Ramsden, personal communication, 1987) indicate that they are incompatible-element-enriched but alkali-poor (i.e., not peralkaline *senso stricto*, but peralkaline in affinity). Thus, the host rocks are similar to some host rocks in the Letitia Lake Group and are similar in geological setting to setting 3.

Table 1 lists the minimum and maximum values of some rare-metal elements in the various host rocks of rare-metal deposits and showings in Labrador. These values have been presented separately for volcanic and intrusive rocks for comparative purposes. Note that a significant rare-metal deposit has been outlined in the Strange Lake peralkaline granite (Dawe, 1984) and a sizeable resource is suggested at the Mann #1, one of the Letitia Lake area syenites (Dujardin, 1960), and within the Letitia Lake Group volcanic rocks. (This compilation of geochemical data should be useful to exploration geologists wishing to evaluate the potential of a particular rare-metal exploration target (i.e., potential of a high-level peralkaline suite or complex)). Thus, targets with rare-metal values within or near the ranges indicated in Table

Table 2. Rare-metal concentrations in deposits and showings in Labrador.

Setting	Deposit or Showing	Zr	Y	Nb	Be	U
Setting 1	Strange Lake granite	5,000	2,000	1,000	200	50
		27,000	16,000	15,000	5,000	300
Setting 2	Flowers River granite	20,000	2,200	1,000	-	-
Setting 3	Letitia Lake showings	40	150	100	30	5
		180	2,400	10,000	2,500	180
Setting 4	Red Wine nepheline syenite	20,000	2,200	1,000	-	-

Sources of data: Strange Lake granite—Miller, 1986; Letitia Lake showings—Miller, 1987; Flowers River granite—Hill, 1981; Red Wine nepheline syenite—Curtis and Currie, 1981; rare-metal values in p.p.m.

Table 3. Rare-metal exploration targets in insular Newfoundland.

Setting	Area	Plutonic Phase	Associated Volcanics
Setting 1, 2	Grand Lake	Hare Hills peralkaline granite	None Identified
Setting 1	La Scie	Seal Island Bight peralkaline granite	None Identified
Setting 1, 3	Green Bay	Kings Point complex (Per±)	Kings Point complex (Per±)
Setting 2, 3	Topsails	Topsails peralkaline granites	Sheffield group (Per±)
Setting 1, 3	Terra Nova	Louil Hills peralkaline granite	Bull Arm Formation (Per?)
Setting 1, 3	Long Harbour	Cross Hills peralkaline granite	Mooring Cove Formation (Per?)
Setting 1, 3	Burin Peninsula	St. Lawrence peralkaline granite	Rocky Ridge Complex
Setting 1, 3	Burin Peninsula	Grand Beach complex (Per?)	Clancey's Pond complex (Per?)

Per? = Peralkaline nature not well established; may have peralkaline trace-element characteristics but does not have peralkaline major-element characteristics. Per± = Definitely peralkaline at some outcrops; not peralkaline or questionably peralkaline in others.

1 should be given further consideration, particularly if the settings are the same.

The ranges of rare-metal values in the showings and deposits in Labrador are listed in Table 2. Not all showings have uniformly high values of all rare-metal elements; e.g., the Letitia Lake showings have abnormally low Zr values but significantly higher Be and Nb values.

Insular Newfoundland

The rock types observed in the settings for rare-metal mineralization described previously for Labrador also occur in insular Newfoundland. Rare-metal targets can be outlined and obviously, any occurrence of felsic peralkaline rocks is a potential target. A preliminary literature survey of anorogenic and peralkaline volcanics and intrusions suggests that the complexes or suites listed in Table 3 are potential rare-metal targets. Each of these potential targets is discussed further.

Hare Hills peralkaline granite—quartz syenite (Figure 3): Occurrences of peralkaline granite, peralkaline granite gneiss and peralkaline quartz syenite in the Grand Lake area, were recently described by Whalen and Currie (1984). However, little is known about this target area, since chemical analyses and detailed geological data have not as yet been published. The available data indicate that any rare-metal mineralization

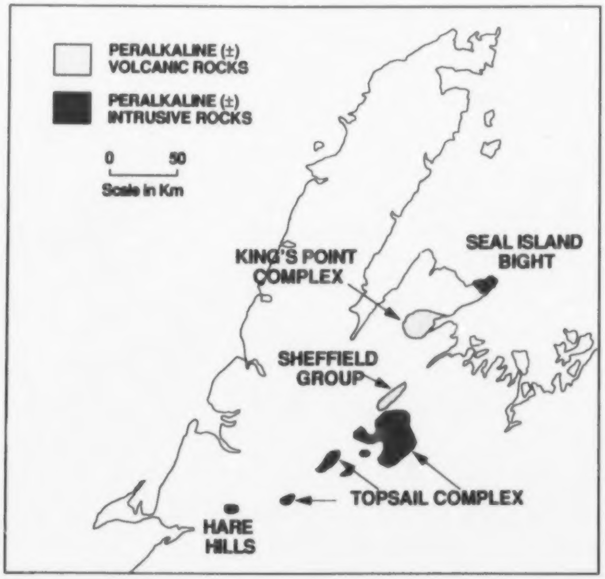

Figure 3. *Rare-metal targets in north-western Newfoundland; peralkaline and peralkaline-related volcanic and high-level plutonic rocks.*

Table 4. Minimum and maximum values of rare-metal concentrations in suggested targets in Newfoundland.

Setting	Host	Zr	Y	Nb	Be	U
Setting 1	Seal Island Bight granite	515	80	33	-	4
		740	-	60	-	-
Setting 2, 3	Kings Point complex	400	60	-	-	-
		800	120	-	-	-
Setting 1, 2	Topsails peralkaline granite	390	50	20	-	3
		1540	340	110	-	9
Setting 2, 3	Sheffield group	805	100	30	-	11
		1660	170	60	-	-
Setting 2, 3	Bull Arm Formation	600	90	100	-	-
		1970	570	490	-	-
Setting 1, 2	Cross Hills granite	220	32	9	4.1	0.8
		1058	268	121	12.7	11.6
Setting 2, 3	Mooring Cove Formation	710	130	30	-	-
		1100	210	80	-	-
Setting 1, 2	St. Lawrence Granite	340	150	100	-	11
		870	-	-	-	-

Sources of data: Seal Island Bight–Whalen *et al.*, 1986; DeGrace *et al.*, 1976; Kings Point Complex–from diagrams in Kontak and Strong, 1986; Topsail peralkaline granite–Whalen *et al.*, 1986; Taylor *et al.*, 1980; Sheffield group–Whalen, 1986; Taylor *et al.*, 1980; Bull Arm Formation–S.J. O'Brien, personal communication, 1988; Cross Hill peralkaline granite–Saunders and Tuach, *this volume*; Mooring Cove Formation–S.J. O'Brien, personal communication, 1988; St. Lawrence granite–Teng and Strong, 1976; Whalen *et al.*, 1986.

associated with these rocks would be in setting 1, or possibly an unvented equivalent of setting 3. This target area requires basic lithogeochemistry and prospecting reconnaissance work to obtain sufficient data to further evaluate it, for any possible exploration program.

Topsail Complex (Figure 3): The Topsail Complex or igneous terrane, described by Whalen and Currie (1983), contains several peralkaline granite intrusions, and peralkaline rhyolitic porphyries and flows. Rare-metal mineralization in this terrrane would be found in setting 1 and/or setting 2. Chemical analyses of several of the peralkaline granites and volcanic units of this extensive terrane (*circa* 3000 km²) have been published by Taylor *et al.*, 1980; Whalen *et al.*, 1986; and Whalen, 1986. Rare-metal concentration in unmineralized samples of granites and volcanic rocks are listed in Table 4. Comparison of these values to those of the Labrador peralkaline granites and peralkaline volcanics (Table 1) clearly show that the Topsail granite rocks fall within the ranges of values, for those from Labrador. The great extent of peralkaline granites in the Topsail terrane and the relatively little data available for the associated volcanic rocks (Sheffield volcanics of Coyle *et al.*, 1986), require an exploration program in this target area, which would include lithogeochemical, prospecting and regional geochemical surveys, before definite detailed follow-up targets can be located. The conclusions of Coyle *et al.* (1986) and Strong and Coyle (1988) suggest that the peralkaline nature of the Sheffield volcanic rocks is due to metasomatic processes rather than primary igneous processes; if this conclusion is true, then these volcanics should be evaluated with a fifth, metasomatic, setting in mind, otherwise setting 2 is most likely.

Kings Point complex (Figure 3): This complex (*circa* 350 km²) consists of alkali to peralkaline syenites and granites and associated peralkaline and non-peralkaline rhyolitic pyroclastics and quartz–feldspar porphyries (Kontak and Strong, 1986; Mercer *et al.*, 1985). Kontak and Strong (1986) suggest that all of these rocks form a subvolcanic–volcanic complex dominated by a cauldron subsidence structure. The geology indicates that rare-metal mineralization would occur within subvolcanic veins and dykes or disseminated in volcanic units similar to setting 2 (Flowers River Igneous Suite) or 3 (Mann #1 Showing). The Kings Point complex closely resembles both the Flowers River and Mann #1 settings in many aspects, including the structural setting and the occurrence of both peralkaline and non-peralkaline rhyolitic volcanic units. Very little geochemical data is available for the peralkaline rocks of this complex (Table 4; trace-element data taken from graphs in Kontak and Strong, 1986; major-element data unavailable), hence little can be said about the potential for rare-metal mineralization. The size and geology of this complex requires an extensive lithogeochemistry and prospecting program, to adequately evaluate the rare-metal potential; an initial lake sediment survey would be valuable in reducing the area to be covered by ground programs.

Seal Island Bight granite (Figure 3): This peralkaline granite (*circa* 4 km²) was mapped as part of the La Scie intrusive suite (DeGrace *et al.*, 1976) on the Baie Verte

Penninsula. The lack of evidence to suggest that this granite was vented to the surface (e.g., absence of associated peralkaline rhyolitic volcanics in the nearby Cape St. John volcanic rocks) indicates that any rare-metal mineralization associated with this intrusion would be found in setting 1. Several chemical analyses have been reported (DeGrace *et al.*, 1976; Whalen *et al.*, 1986; Table 4). The comparison of these data with those of similar peralkaline granites in Labrador (Table 1) indicates that the Seal Island Bight granite is a poor rare-metal exploration target. A thorough island-wide exploration program should include basic lithogeochemistry and prospecting reconnaisance to verify and support this conclusion and the earlier published geochemical data.

Cross Hills Plutonic Suite (Figure 4): The Cross Hills Plutonic Suite contains a peralkaline granite (*circa* 8 km^2), which has aplite veins containing significant values of Zr, Y and Nb (Hopfengartner, 1982; Tuach, 1984; Saunders and Tuach, *this volume*) similar to setting 1 or setting 2. The regional geology (O'Brien *et al.*, 1984) indicates that the spatially related Mooring Cove Formation felsic volcanic rocks (Long Harbour Group) may be extrusive equivalents of or otherwise related to the Cross Hills Plutonic Suite; i.e., related in a manner similar to that of the Letitia Lake Group non-peralkaline volcanic rocks and peralkaline intrusives. Rare-metal ranges for the Cross Hills peralkaline granite (Strong *et al.*, 1974) and the Mooring Cove rhyolitic rocks (S.J. O'Brien, personal communications, 1988) are in the lower portion of the ranges for the Labrador peralkaline rocks. Yet, the occurrence of high values (Saunders and Tuach, *this volume)* in veins and dykes indicate that the Cross Hills peralkaline granite contains good potential for significant rare-metal mineralization. An exploration program of the peralkaline granite would require a detailed prospecting and lithogeochemistry survey. A program for the Mooring Cove Formation would initially require a regional lake sediment or lithogeochemical survey to reduce the target area, and to follow this up, with detailed lithogeochemical and prospecting surveys.

Louil Hills peralkaline granite (Figure 4): This peralkaline granite (*circa* 5 km^2) is part of the Louil Hills suite, which was mapped and described recently by O'Brien (1987). The interpretation of the regional geology (O'Brien *et al.*, 1988) indicates that the Bull Arm Formation (Musgravetown Group) is spatially, chemically and temporally related to the Louil Hills suite, in a manner similar to that of the Mooring Cove Formation and the Cross Hills Plutonic Suite (i.e., the Bull Arm Formation is not peralkaline in major-element chemistry, but has similar trace-element characteristics). Rare-metal mineralization in the Louil Hills area, if present, would probably be found in either settings 1 or 2. Available chemical analyses for the Bull Arm Formation rhyolitic rocks (Table 4; S.J. O'Brien, personal communication, 1988), when compared with analyses of felsic volcanics associated with rare-metal mineralization (Table 1), indicate that there is good potential for volcanic-hosted mineralization. These rocks have the best Y–Nb potential of any volcanic rocks sampled in insular Newfoundland. Evaluation of the rare-metal potential of the Louil Hills peralkaline granite could be accomplished with a relatively minor lithogeochemical and prospecting program, due to the limited areal extent of these intrusive rocks. The regional occurrence of the Bull Arm Formation (> 500 km^2) would require a major lake sediment geochemical program, to locate follow-up targets for an extensive lithogeochemistry and prospecting program.

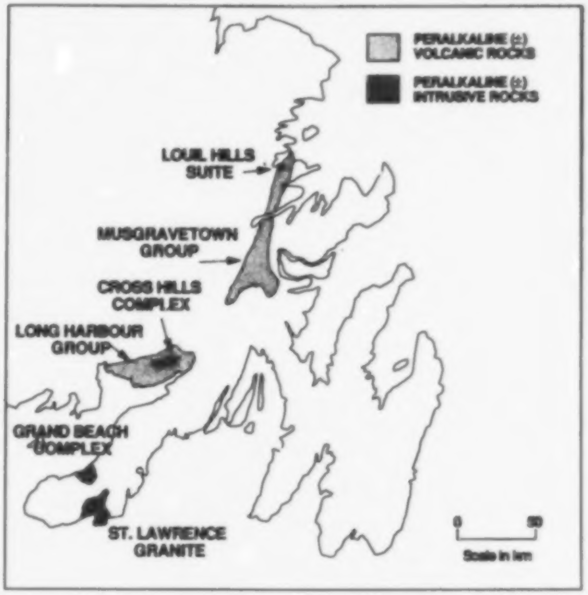

Figure 4. *Rare-metal targets in south–eastern Newfoundland; peralkaline and peralkaline-related volcanic and high-level plutonic rocks.*

St. Lawrence Granite (Figure 4): This peralkaline granite (*circa* 40 km^2) is well known as an economic source of fluorite, which occurs in numerous veins. It also has good potential for rare-metal mineralization in either settings 1 or 2 veins, or disseminated in setting 2 peralkaline volcanics. The St. Lawrence peralkaline granite has been the focus of extensive exploration for fluorite veins; significant rare-metal mineralization was not located during any of this exploration activity. Mapping in the surrounding region has uncovered peralkaline rhyolitic flows and ignimbrite (Rocky Ridge complex; Strong *et al.*, 1978) and peralkaline quartz-feldspar porphyries, which may be extrusive equivalents of the granite. These possible extrusive equivalents have not been sampled or evaluated for rare-metal mineralization and have unknown potential. Chemical values for rare-metal elements (Teng and Strong, 1976; Whalen *et al.*, 1986) from the St. Lawrence Granite (Table 4) indicate a low potential for significant mineralization. Reported analyses for some rare metals from the fluorite veins (Strong *et al.*, 1984) indicate that Y values are significant in the fluorite. This suggests that all of the known fluorite veins should be sampled for rare-metal mineralization. It also indicates that the granite has good potential for mineralization in other kinds of veins, if they are present.

Grand Beach complex (Figure 4): The rocks of this complex are (Strong *et al.*, 1978) possible equivalents of the St. Lawrence Granite as they have similar trace-element characteristics. Chemical analyses for these rocks have, as yet, not been published, but unpublished results (see Table 4; S.J. O'Brien, personal communication, 1988) indicate that although these rocks do not have Agpaitic Indices greater than one, they do have the high trace-element values indicative of rocks of peralkaline affinity; however, peralkaline minerals have not been observed. Mapping and core-logging work (Pitman, 1978) has indicated that most of these rocks are rhyolitic tuffs, ignimbrites and agglomerates. A reconnaissance lithogeochemistry and prospecting program would be required to evaluate their suitability as a target, and a further analysis of the BP core (Pitman, 1978) for rare-metal elements would aid in the evaluation process.

CONCLUSIONS

Metallogenic studies of rare-metal mineralization in Labrador has resulted in the development of a metallogenic model for high-level felsic peralkaline rock-hosted rare-metal mineralization. This model can be applied to insular Newfoundland to assist geologists when exploring for rare-metal deposits.

A literature survey of the geology of insular Newfoundland indicates that there are eight occurrences of extrusive and/or intrusive rocks of peralkaline nature or peralkaline affinity and these occurrences can be ranked as to their exploration potential as follows:

1) Louil Hills peralkaline granite–Bull Arm Formation–The high values of Y (570 ppm) and Nb (490 ppm) make the Bull Arm Formation a very favourable exploration target.

2) Cross Hills peralkaline granite–Several aplite veins contain high grades, but insignificant tonnages of rare-metal elements, indicate that rare-metal concentration processes were operative during the formation of this granite, however, it is difficult to evaluate if significant recoverable tonnages were produced.

3) Topsails peralkaline granite–The available analyses for the numerous granite bodies in the Topsails igneous terrane are not encouraging, but the large number and great areal extent of these bodies indicate that much more work is required, before these peralkaline granites can be considered barren. The related Sheffield volcanic rocks also have an unknown potential.

4) St. Lawrence peralkaline granite–High values of Y in fluorite veins indicate that rare-metal element concentration processes were operative during the formation of this granite. Evaluation of rare-metal concentrations in all fluorite veins has not been attempted and a search for rare-metal-enriched pegmatite–aplite veins has not been carried out.

5) Kings Point complex–The large areal extent and lack of rare-metal analyses make the rare metal potential of the rocks of this complex unknown; further work is needed to better evaluate its potential.

6) Grand Beach complex–The lack of extensive analyses of the rocks of this complex make its potential unknown.

7) Seal Island Bight peralkaline granite–The relatively poor concentration of rare-metal elements in this granite indicate that it is a low priority target; reconnaissance work is needed to better evaluate its potential.

8) Hare Hills peralkaline rocks–The lack of data and small areal extent of these rocks make them a low priority target; more information is required to better evaluate these rocks.

This preliminary evaluation indicate that there are several high-potential exploration targets in insular Newfoundland.

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Zr-Nb-Y-REE MINERALIZATION IN THE CROSS HILLS PLUTONIC SUITE

C.M. Saunders and J. Tuach[1]
Mineral Deposits Section

ABSTRACT

Minor Zr-Nb-Y-REE mineralization occurs in peralkaline granitic rocks of the Cross Hills Plutonic Suite. The suite consists of spatially and genetically related gabbro to diabase, granodiorite, biotite granite, peralkaline granite and minor syenite intrusions, which outcrop over 260 km² in the Terrenceville map area of southeast Newfoundland.

Euhedral and botryoidal, zoned, hydrothermal zircon is disseminated in narrow aplite veins and concentrated in tuffisite and along fractures in the peralkaline granite. The mineralized samples are significantly enriched in Zr, Y, Nb, Sn, Th, Ce and heavy REE. Assays have returned values of up to 20,000 g/t Zr, 2968 g/t Y, 2334 g/t Nb, 1550 g/t Ce, 725 g/t Th and 240 g/t Sn.

Scanning electron microscopy reveals a variety of Zr-, Nb-, Y- and REE-bearing minerals associated with the zircon. Baddelyite, fergusonite and thorite have been tentatively identified, along with thorium oxide and additional unidentified minerals.

The mineralogy of the occurrences and the distribution of mineralization in aplite dykes, tuffisite and fractures indicates a magmatic–hydrothermal origin. The peralkaline granite provided a source for the mineralizing elements. Concentration and mobilization of these elements was aided by the prevalence of oxidizing, alkaline and possibly carbonate-rich fluids.

INTRODUCTION

The Cross Hills Plutonic Suite (Bradley, 1962; Tuach, 1984, *in preparation;* O'Brien *et al.*, 1984) consists of a group of spatially and genetically related gabbro to diabase, granodiorite, biotite granite, peralkaline granite and minor syenite intrusions, which outcrop over 260 km² in the Terrenceville (1M/10) map area of southeast Newfoundland (Figure 1). In 1983, mapping and sampling was focused on an area of about 20 km² to the northwest of Grand Le Pierre (Figure 2). The peralkaline granite in this area is host to minor Zr-Nb-Y-REE mineralization.

Preliminary results of field work were presented in Tuach (1984) and the first draft of a report, documenting major and trace element data, petrographic observations and geological interpretations has been completed (Tuach, *in preparation)*. This present paper summarizes the geology, geochemistry and mineralization of the Cross Hills Plutonic Suite and provides details of scanning electron microscopy (SEM) of mineralized samples and results of REE analyses of six peralkaline granite samples.

GENERAL GEOLOGY

The following is summarized from Tuach (1984 and *in preparation).* The high-level, posttectonic Cross Hills Plutonic Suite intrudes late Hadrynian, subaerial, low-grade, metavolcanic rocks (Belle Bay Formation) and metasedimentary rocks (Anderson's Cove Formation) of the Long Harbour Group (Williams, 1971; O'Brien and Nunn, 1980; O'Brien *et al.*, 1984). The Belle Bay and Anderson's

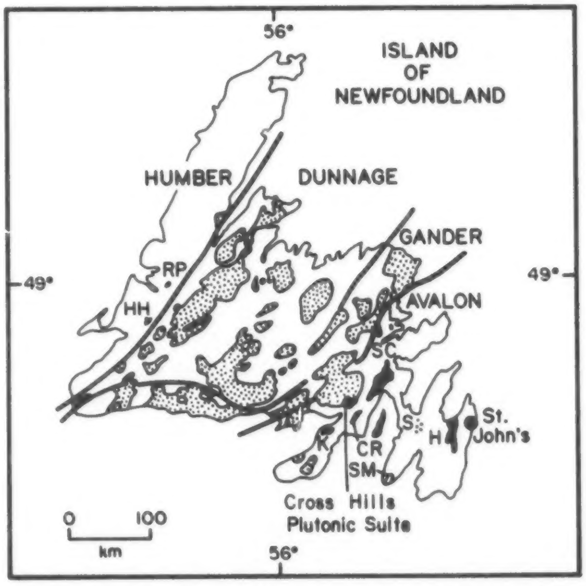

Figure 1. *Location of the Cross Hills Plutonic Suite in Newfoundland; also shown are the major tectonostratigraphic subdivisions in Newfoundland, and location and distribution of Paleozoic (stippled) and Eocambrian to Cambrian plutonic rocks (black). RP–Round Pond, HH–Hare Hill, L–Louil Hills, SC–Swift Current, CR–Cape Roger Mountain, K–Knee Granites, S–Spreadeagle, SM–Saint Marys Diabase, H–Holyrood.*

This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

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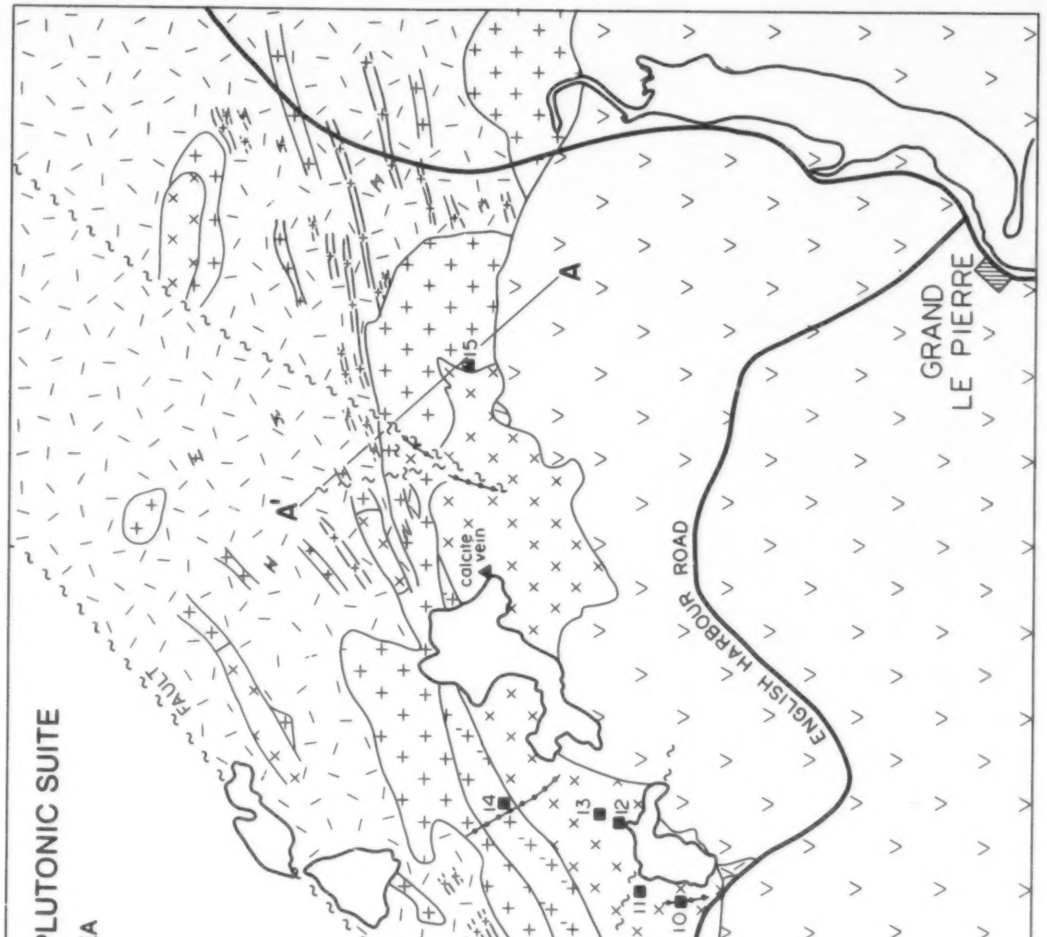
GEOLOGY OF THE CROSS HILLS PLUTONIC SUITE
GRAND LE PIERRE AREA
CAMBRIAN
CROSS HILLS PLUTONIC SUITE
× Peralkaline Granite
+ Biotite Granite; +' Aplite
\ Diabase – Gabbro
PRECAMBRIAN
BELLE BAY FORMATION
> Predominantly silicic volcanic rocks
■ Mineral Occurrence
Quartz Vein
0
500m
N
CROSS HILLS FAULT
ENGLISH HARBOUR ROAD
GRAND LE PIERRE
calcite vein
A
A'
B
B'
Pond
Bart
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15

Figure 2. *Geology of the Cross Hills Plutonic Suite in the Grand Le Pierre area.*

Cove formations are possibly facies-equivalent and are unconformably overlain by peralkaline ash flows of the Mooring Cove Formation. The latter has been interpreted as the extrusive equivalent of the peralkaline granite phase of the Cross Hills Plutonic Suite (O'Brien *et al.*, *in press)*. The Early Cambrian Cross Hills Plutonic Suite has provided imprecise Rb-Sr ages of around 560 ± 30 Ma (Tuach, *in preparation)*. The northern boundary of the suite is cut by the Sage Pond phase of the Ackley Granite suite. The Sage Pond phase has provided ^{40}Ar/^{39}Ar ages between 368.4 ± 4.2 Ma and 371.4 ± 5.4 Ma (Kontak *et al.*, 1988).

The Cross Hills Plutonic Suite is one of several Eocambrian to Cambrian plutonic bodies, commonly of alkaline to peralkaline affinity, which are found in both the Humber and Avalon zones of Newfoundland (Figure 1). It has been divided into five map units on the basis of field and geochemical data (Table 1). These are: 1) a mafic unit consisting of gabbro and diabase; 2) granodiorite; 3) biotite granite; 4) peralkaline granite, and 5) minor syenite. The following descriptions are summarized from Tuach *(in preparation)*.

Mafic unit

The mafic unit contains a variety of dark green, to grey, to black gabbro, diabase and minor dioritic rocks that are intruded by numerous veins and dykes of biotite granite. Small roof pendants and xenoliths of silicic volcanic rocks occur locally, and are common to the east of Grand Le Pierre. Gabbro is medium- to coarse-grained with local pegmatite patches. Diabase is fine- to medium-grained with an ophitic texture. Fine-grained aphyric and plagioclase- and pyroxene-phyric varieties of diabase also occur. Xenoliths of earlier gabbroic and diabase phases are seen in some exposures of the mafic unit. Diorite is the most abundant lithology at the Southeast Bight Hills and at Nine Mile Hill (O'Brien and Nunn, 1980).

Granodiorite

Mottled, purple to orange, massive, equigranular, medium- to coarse-grained hornblende–biotite granodiorite outcrops over a large area to the east of Grand Le Pierre. Abundant felsic and mafic volcanic and diabase xenoliths are common. The granodiorite is locally cut by diabase and aplite dykes, which are up to a metre wide.

Biotite granite

This is an orange, miarolitic, equigranular granite containing up to 5 percent mafic minerals. Although most of the granite is fine- to medium-grained, it is characterized by zones of aplitic texture, and an important mappable subunit of aplite of similar mineralogy and composition, occurs along the southern boundary of the biotite granite to the northwest of Grand Le Pierre (Figures 2 and 3). Locally, fine- to coarse-grained pegmatite is present. Related, thin, felsic granite dykes that locally intrude the mafic rocks, grade from an aplitic margin to a pegmatitic core. Xenoliths of felsic volcanic and mafic intrusive rocks are common in exposures to the northeast of Grand Le Pierre. North and west of Grand Le Pierre, silicic volcanic xenoliths are rare, and xenoliths of mafic intrusive rocks occur only in close proximity to contacts with the main bodies of mafic rocks. Epidote veins and fractures are common, particularly near faults and lineaments. Networks of quartz-filled fractures and veins are locally common, and quartz veins occur up to one metre wide and 100-m long.

Peralkaline granite

The peralkaline granite is purple-brown, miarolitic, fine- to medium-grained and is characterized by Na-amphibole, which is partly to completely replaced by chlorite. Granophyric intergrowths can commonly be recognized in hand-specimens, and small pegmatite patches occur locally. Tuffisite, indicative of gas streaming, is common and small peralkaline aplite veins occur. Several small, mineralized, peralkaline aplite veins intrude peralkaline granite at mineral occurrence 12 (Figure 2, Table 2). Hematite staining is common, particularly around fractures and tuffisite. Clay alteration occurs in association with fractures near the southern contact between the granite and the silicic volcanic rocks. Quartz vein networks and quartz veins up to a metre wide occur and are locally continuous with those in the biotite granite.

Syenite

A 50-m-long outcrop of purple, manganese-stained Na-amphibole-bearing syenite was found on the Grand Le Pierre road and a small outcrop of orange, alkali feldspar granite was noted 1.5 km to the south. These rocks are medium- to coarse-grained and equigranular. Two 50-m-wide, east-northeast-trending, subvertical dykes of fine-grained, red–brown, quartz-poor trachyte, outcrop approximately 1 km to the west of the main exposure of the Cross Hills Plutonic Suite. They exhibit high-background radioactivity and are tentatively correlated with the alkalic and peralkaline phase of the Cross Hills Plutonic Suite.

MINERALIZATION AND ALTERATION

The peralkaline granite and immediately adjacent rocks are host to numerous small mineral occurrences, many of which were located by Saarberg Interplan (Hopfengaertner, 1982). These are described by Tuach *(in preparation)* and are known as occurrences 1 to 15 (Figure 2, Table 2).

The most significant occurrences (11 and 12) are anomalous concentrations of Zr-Nb-Y-REE minerals. Radioactivity is 4 to 5 times above background levels at the mineralized locations, with spot highs up to 10 times above background levels. Other similar radioactive occurrences are listed in Table 2.

Table 1. Summary of geochemical data from the Cross Hills Plutonic Suite, Sage Pond–Grand Le Pierre area.

	GABBRO <52% SiO_2		BIOTITE GRANITE		APLITE		GRANODIORITE*		PERALKALINE GRANITE		SYENITE		MINERALIZED PERALKALINE	
	$\bar{x}$	$\bar{s}$	$\bar{x}$	$\bar{s}$	$\bar{x}$	$\bar{s}$	$\bar{x}$	$\bar{s}$	$\bar{x}$	$\bar{s}$	$\bar{x}$	$\bar{s}$	JT-067	JT-124
SiO_2	47.9	2.78	74.4	1.22	75.1	0.56	69.4	0.92	76.6	0.54	65.1	2.55	74.3	73.8
Al_2O_3	16.3	1.47	12.6	0.38	12.4	0.19	14.3	0.23	11.0	0.37	16.1	0.72	8.75	8.00
Fe_2O_3	3.64	1.74	0.73	0.14	0.68	0.15	1.08	0.22	1.88	0.45	2.38	0.93	6.14	5.80
FeO	6.96	1.70	0.76	0.26	0.70	0.18	1.67	0.25	0.45	0.18	2.36	1.08	0.41	0.11
MgO	7.23	1.26	0.26	0.16	0.15	0.04	0.79	0.16	0.05	0.03	0.11	0.08	0.16	0.08
CaO	9.02	1.08	0.62	0.47	0.45	0.06	2.59	0.21	0.09	0.07	0.44	0.27	0.06	0.13
Na_2O	2.85	0.42	3.70	0.22	3.89	0.18	4.06	0.12	3.75	0.21	6.01	0.58	1.65	1.75
K_2O	1.07	0.65	5.16	0.40	5.16	0.26	2.63	0.16	4.57	0.29	5.14	0.79	5.15	4.62
TiO_2	1.75	1.01	0.28	0.05	0.25	0.04	0.36	0.01	0.18	0.03	0.37	0.21	0.36	0.57
MnO	0.19	0.06	0.03	0.01	0.03	0.01	0.09	0.01	0.04	0.03	0.13	0.06	0.06	0.19
P_2O_5	0.29	0.18	0.03	0.01	0.02	0.01	0.06	0.01	0.02	0.01	0.05	0.04	0.02	0.03
LOI	2.30	0.64	0.71	0.13	0.54	0.11	1.28	0.18	0.49	0.11	1.07	0.42	0.93	0.79
CO_2	0.072	0.042	0.072	0.034	0.047	0.027	0.127	0.107	0.075	0.072	0.128	0.133	-	-
Li	36	9.7	11	2.8	8.1	1.9	13	2.5	8.8	6.5	13	4.8	10	8
Be	1.8	0.5	5.7	5.5	5.3	0.6	1.9	0.1	8.5	2.2	7.8	7.7	43.8	51.9
F	449	254	165	77	394	246	-	-	79	71	1196	1427	54	141
V	276	129	16	6.5	11	2.2	38	7.5	6	1.9	6.8	5.9	10	15
Cr	136	36	1.9	2.4	0.6	1.1	2.7	2.1	2	2.2	0.3	0.8	0	0
Ni	72	39	1.7	0.08	1.2	0.5	1	0	1	0	1	0	-	-
Cu	62	23	3.8	0.07	3.8	0.8	8.7	5	12	16	3	1	-	-
Zn	99	27	50	19	54	23	51	4.6	115	86	147	78	149	162
Ga	20	5	20	2.4	22	1.19	16	1.5	27	1.8	35	7	26	11
Rb	51	30	175	22	167	8.4	86	4.7	193	21	141	43	309	279
Sr	352	108	47	26	33	5.5	180	8.5	11	5.7	38	27	0	4
Y	35	13.7	83	27.5	97	12	31	4.4	150	59	123	74	1619	2968
Zr	136	68	231	19	259	28	154	3.1	690	234	996	612	6800	19208
Nb	9.6	2.9	30	5.3	35	2.7	13	0.6	65	28	244	176	809	1860
Ag	0.1	0	0.1	0	0.1	0	0.1	0	0.1	0.1	0.1	0.1	0.1	0.1
Sn	1	-	1	0	2	1.6	-	-	5.2	3	7	5.2	40	90
Ba	164	173	307	84	248	88	647	62	73	64	387	377	47	220
La	11.7	6	46	13	62	15	18	2.3	34	22	124	95	33	190
Ce	109	51	194	41	222	29	84	13	200	67	351	224	762	1550
Pb	2.7	2.5	13	4.4	15	9.1	7.3	2.5	18	13	18	22	33	44
Th	2.8	3.4	25	2.9	27	2.9	13	1	31	13	32	23	364	725
U	0	-	4.1	1	4.8	0.6	-	-	6.2	2.7	6.5	3.7	64.2	112.5
n**	12		20		9		3		22		6		1	1

* These samples are from the Cross Hills Plutonic Suite–east of Grand Le Pierre.

** Maximum number of samples. For some elements, i.e., F, Ni, Cu, Sn and U, n may be lower.

$\bar{x}$–mean; $\bar{s}$–standard deviation; major elements and CO_2 in percent, trace elements in g/t; (-) not analyzed

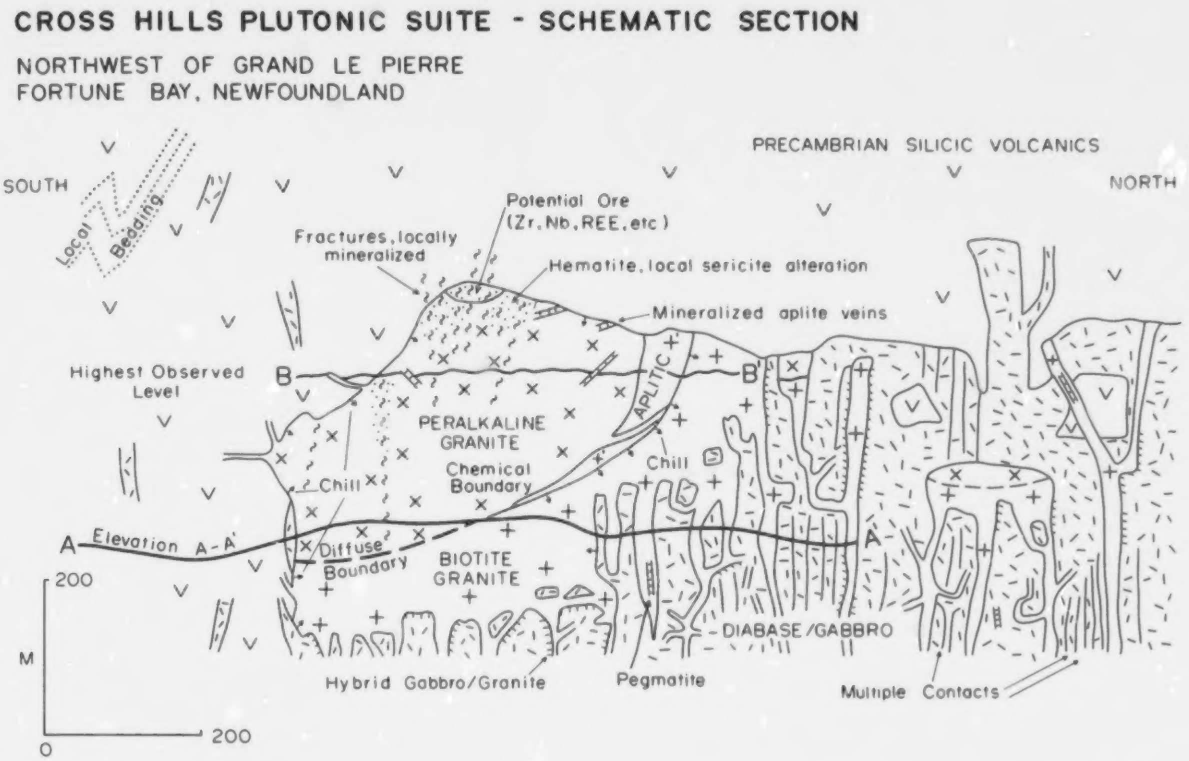
CROSS HILLS PLUTONIC SUITE - SCHEMATIC SECTION
NORTHWEST OF GRAND LE PIERRE
FORTUNE BAY, NEWFOUNDLAND
PRECAMBRIAN SILICIC VOLCANICS
SOUTH
NORTH
Local Bedding
Potential Ore (Zr, Nb, REE, etc)
Fractures, locally mineralized
Hematite, local sericite alteration
Mineralized aplite veins
Highest Observed Level
B
B'
APLITIC
PERALKALINE GRANITE
Chemical Boundary
Chill
Chill
A
Elevation A - A'
A'
Diffuse Boundary
BIOTITE GRANITE
DIABASE/GABBRO
Hybrid Gabbro/Granite
Pegmatite
Multiple Contacts
200
M
0
200

Figure 3. *Schematic (composite) geological section across the Cross Hills Plutonic Suite in the area to the northwest of Grand Le Pierre; level of sections shown on Figure 2 and denoted by A-A', B-B'.*

The highest assay results were obtained from chips of small aplite veins that intrude peralkaline granite exposed in a 3- by 2-m-trenched outcrop at occurrence 12 (Plate 1). The outcrop contains 5 to 10 percent, gently-dipping, up to 5-cm-wide, sheeted aplite veins; the host granite has weak anomalous radioactivity and did not give anomalous analytical values. A selected sample of vein material (JT-124) assayed 19208 g/t Zr, 2968 g/t Y, 1860 g/t Nb, 1550 g/t Ce, 725 g/t Th and 90 g/t Sn. Similar values were obtained by Hopfengaertner (1982). Abundant fine-grained (less than 0.1 mm), euhedral zircon is disseminated in the aplite matrix and is concentrated in tuffisite and fractures in the veins. It is commonly intergrown with hematite (Plate 2). SEM analysis reveal a variety of Zr-, Y-, Nb- and REE-bearing minerals. These include traces of baddelyite (ZrO_2) and possible fergusonite ($YNbO_4$) (Figure 4a). Additional unidentified Nb- and REE-bearing minerals were found (e.g., Plate 3 and Figure 4b, c and d).

At occurrence 11 (Figure 2) a 10- by 5-m area of anomalous radioactivity (up to 5 times above background levels) occurs in a south-facing scarp. A selected sample (JT-067) of radioactive chips assayed 6800 g/t Zr, 1619 g/t Y, 809 g/t Nb, 762 g/t Ce, 364 g/t Th and 40 g/t Sn. Botryoidal, zoned zircon (Plate 4) is concentrated in tuffisite and fractures, and fills spaces between feldspar crystals (Plate 5). Zoning may be caused by variations in Fe and Mn content (determined by SEM analysis). Some zircons are cored by thorium silicate (Plate 6 and Figure 4e), which may be thorite ($Th[SiO_2]$). Thorium oxide is also common (Plate 6 and Figure 4f). This is isotropic and may be a metamict alteration product of thorium silicate; unidentified Th-Nb minerals were also found (Figure 4g and h).

The Cross Hills Plutonic Suite is also host to several minor base-metal occurrences. Red-brown, hematite- and limonite-stained peralkaline granite, containing 1 to 5 percent fine-grained disseminated pyrite is ubiquitous over an area of 1000 by 200 m along the English Harbour road; tuffisite zones in the granite contain abundant hematite grains. Traces of galena and up to 3 percent fine-grained disseminated and fracture-filling pyrite are found in subvertical 30- by 3-m zones of pervasive grey-green sericite alteration (occurrence 4). Small areas of malachite stain, in some instances associated with traces of chalcocite and bornite, occur in quartz veins that cut the biotite and peralkaline granites (occurrences 9, 10 and 14) and in sheared rhyolite (occurrence 1). Minor pyrite and galena are found in quartz veins, which cut the peralkaline granite (occurrence 5).

Table 2. Mineral occurrences in and adjacent to the Cross Hills Plutonic Suite. Assays are of selected grab samples.*

Occurrence No.	Mineral or Anomaly	Description**		Zr	Nb	Y	Ce	Th	Sn	Sample No.	Comments	REF.
1	Malachite	Minor stain in sheared rhyolite	Trace over 1 m²								Not analyzed	T
2	Radioactive. Up to 6 x background	Patchy. In isolated fractures in peralkaline granite	Fractures over 100 x 20 m area	581	Tr	799	Tr	620	Tr	JT-131		T
3	Radioactive	In peralkaline granite	-	11661	805	n.a.	n.a.	268	93	CH-3B-1		H
4	Pyrite, galena	Disseminated in subvertical 2-m wide sericitized zone in peralkaline granite	2 percent pyrite, trace galena	1190	Tr	139	Tr	Tr	Tr	JT-129	Also 481 g/t Pb and 840 g/t Zn	T
5	Pyrite, galena	In quartz vein cutting, peralkaline granite	Trace								Not analyzed	H
6	Radioactive	In peralkaline granite	-	9518	556-	n.a.	n.a.	259	70	19		H
7	Radioactive, pyrite	Fractures, weak anomalous radioactivity in peralkaline granite	Up to 5 percent pyrite							JT-86	All values at background levels	T
8	Radioactive	In peralkaline granite	-	7576	974	n.a.	n.a.	477	47	18		H
9	Chalcocite, malachite	Veinlets in 1-m wide quartz vein cutting; biotite granite in fault zone	Trace over 0.5 m² area							JT-57		T
10	Malachite	In 0.5-m wide quartz vein cutting, peralkaline granite	Trace over 0.5 m²								Not analyzed	T
11	Radioactive	Fracture surfaces in south-facing scarp in peralkaline granite	10 m² area	6800	809	1619	762	364	40	JT-67		T
12	Radioactive, hematite	In aplite veins (to 12 cm, average 3 cm wide) in peralkaline granite. Abundant fine-grained zircon in fractures	1 x 2 m area	20000 19208	2334 1860	n.a. 2968	n.a. 1550	550 725	240 90	11 JT-124	Also 108 g/t Ta and 41 g/t U	H T
13	Radioactive	In peralkaline granite	-	10208	425			187	118	14		H
14	Malachite, chalcocite	In 2-m-wide quartz vein cutting biotite granite	Trace								Not analyzed	H
15	Radioactive	-	-	6328	316			188	37	22		H

H – Hopfengaertner (1982); T – Tuach (in prep.); * – All values in g/t. Tr – trace; n.a. – not analyzed; ** – Unavailable or sketchy for some of the Saarberg Interplan occurrences

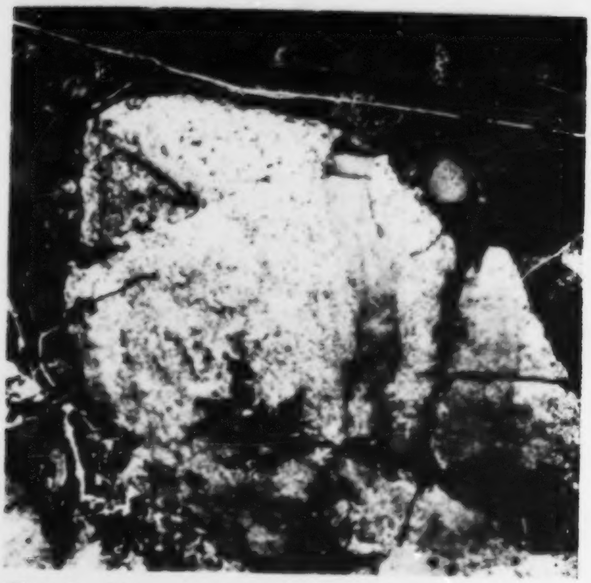

Plate 1. *Mineralized peralkaline aplite veins cut medium-grained peralkaline granite at occurrence 12.*

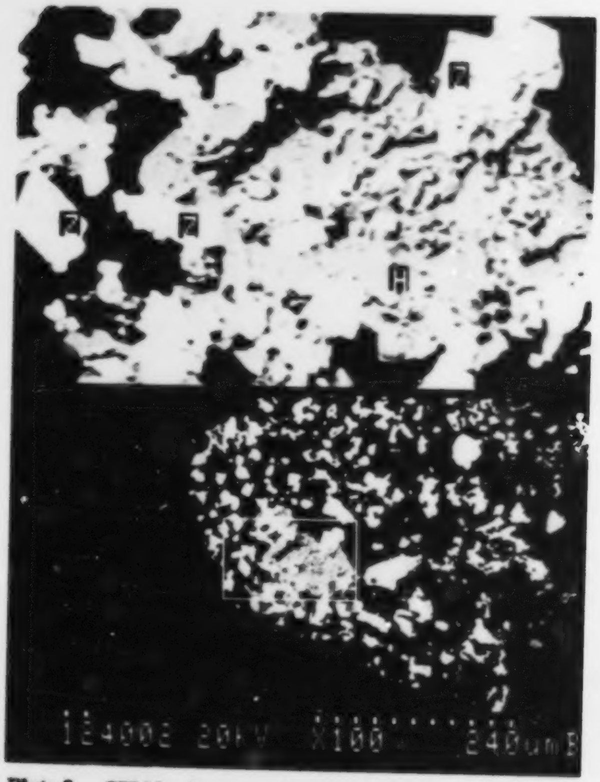

Plate 2. *SEM backscatter photograph of euhedral zircon (Z) intergrown with hematite (H); black is silicate gangue; scale bar refers to lower half of photo; scale of top (area in box) is 5x scale of bottom; sample JT-124.*

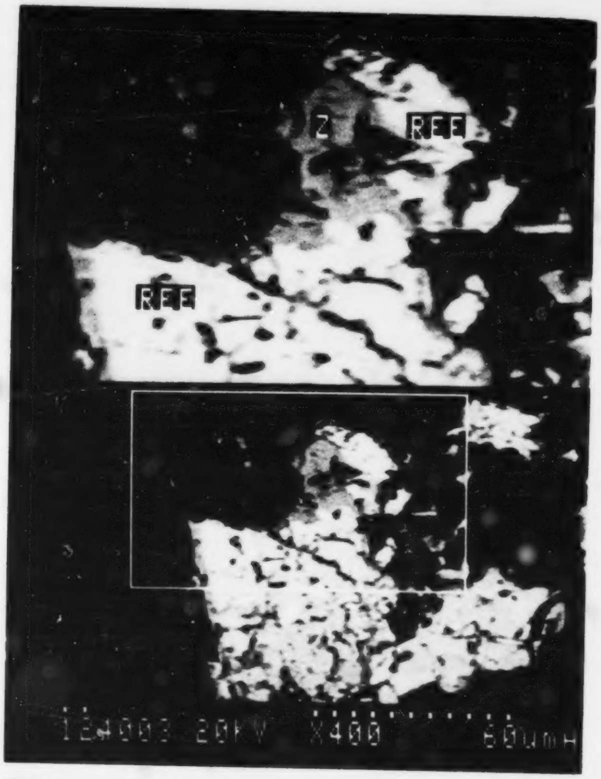

Plate 3. *SEM backscatter photograph of REE mineral intergrown with zircon (Z); spectral scan of REE mineral is shown in Figure 4d; scale of top is 2x scale of bottom; sample JT-124.*

RARE-EARTH ELEMENTS

Six samples of peralkaline granite, including a mineralized sample from occurrence 11 and one from the mineralized peralkaline aplite vein at occurrence 12, were analyzed for rare-earth elements by ICP-MS. The analytical methods are given in Tuach *(in preparation)* and the results are shown in Figure 5. The four unmineralized peralkaline granite samples have relatively flat REE patterns and negative Eu anomalies. However, sample 437 is enriched in both light REE (LREE) and heavy REE (HREE). The two mineralized samples, 067 and 124, are enriched in total REE, but have highly enriched, convex-upwards HREE patterns. All samples except 155 have HREE maxima at Tm. Four samples also have extremely high Ce values that appear to be somewhat independent of HREE concentrations. A Ce-rich mineral has been found in sample 124 (Figure 4d).

DISCUSSION

Significant enrichment of HREE over LREE is a common feature of similar peralkaline granite-hosted mineralization. Taylor *et al.* (1981) and Taylor and Fryer (1983) documented a five to seven-fold HREE enrichment in

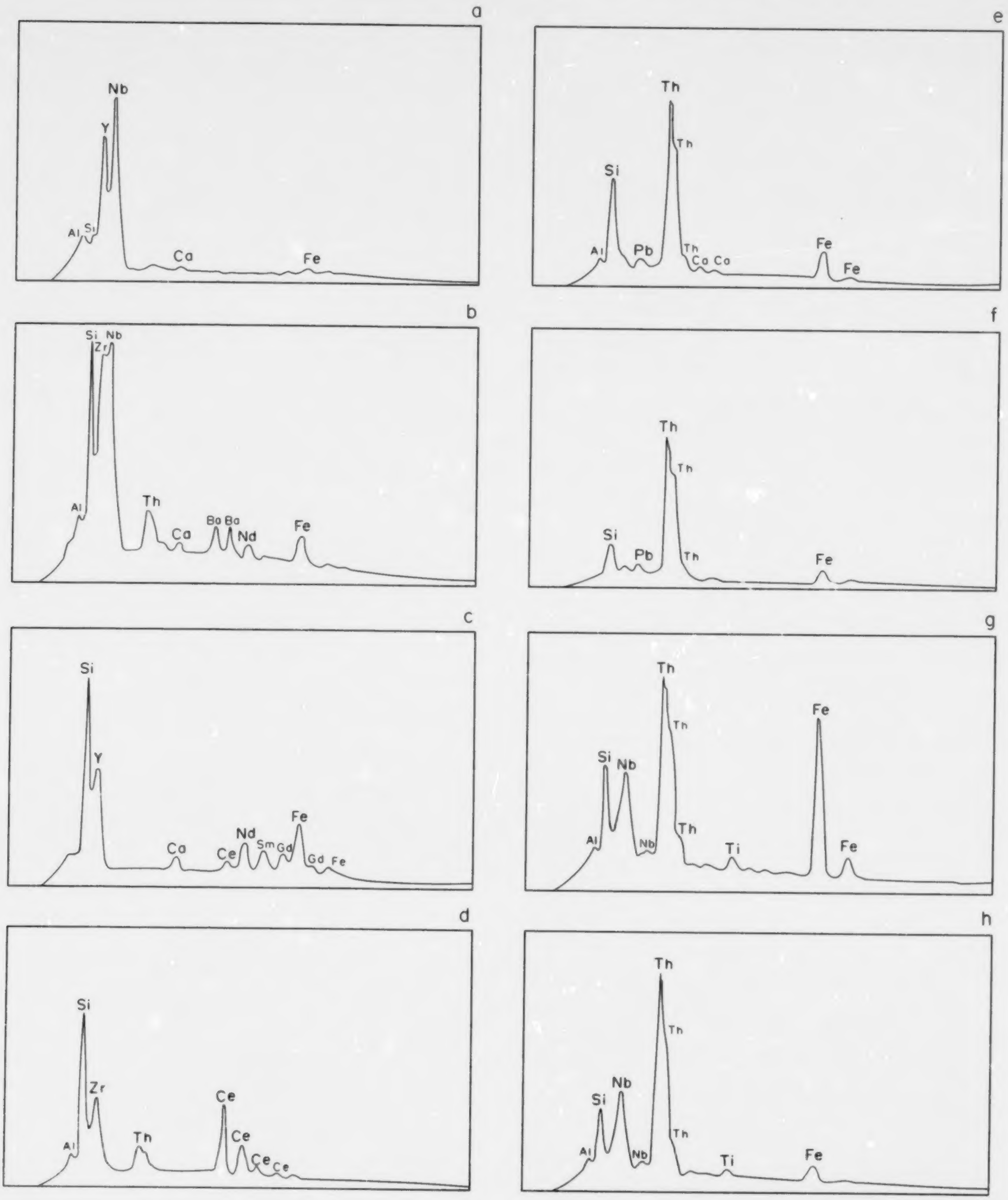

Figure 4. *Scanning electron microscope spectra of minerals from occurrence 11–sample JT-067 and occurrence 12–sample JT-124: a) fergusonite?–sample 124; b,c,d) unidentified minerals–sample 124; e) thorite?–sample 067; f) thorium oxide–sample 067; g,h) unidentified minerals–sample 067.*

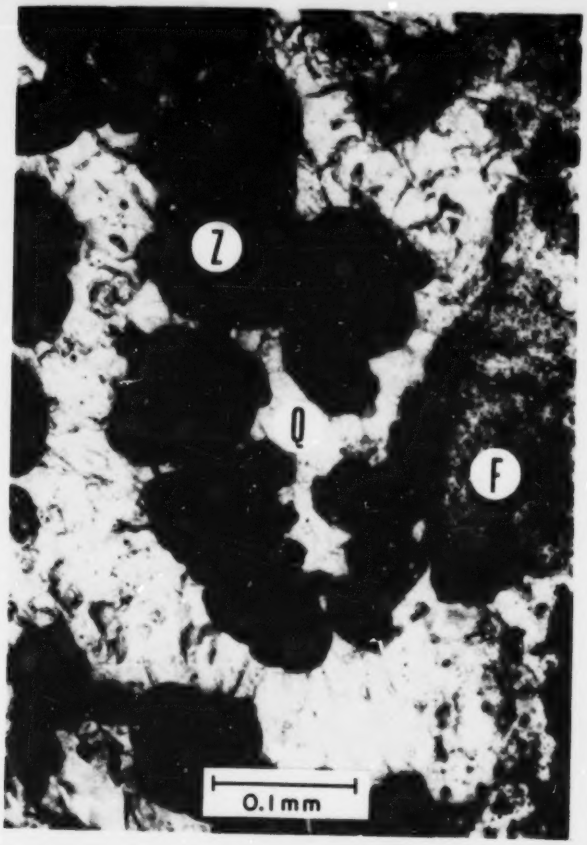

Plate 4. *Photomicrograph of hydrothermal, botryoidal, zoned zircon (Z) from occurrence 11; also visible are K-feldspar (F) and quartz (Q); sample JT-067, plane light.*

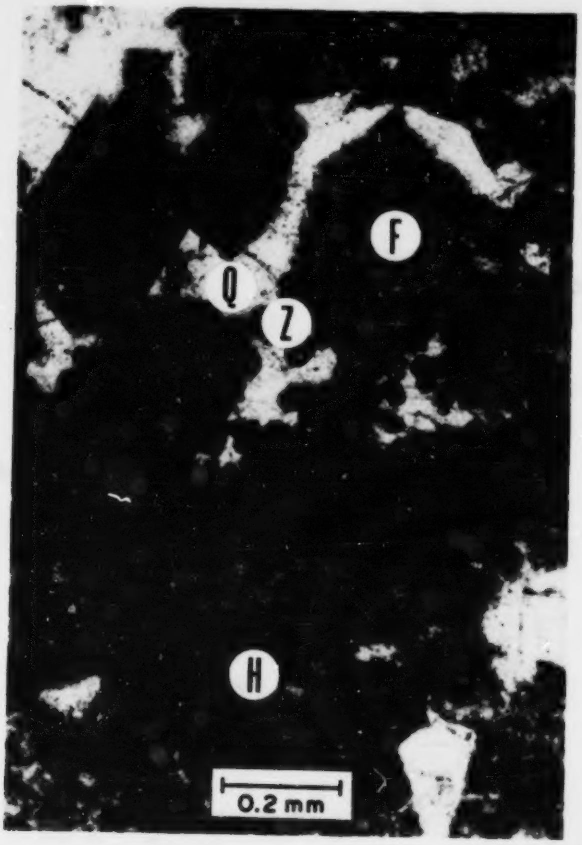

Plate 5. *Photomicrograph of botryoidal zircon (Z) surrounding K-feldspar crystals (F); also visible are quartz (Q) and fine-grained hematite (H); sample JT-067, plane light.*

peralkaline rocks of the Topsails Igneous Complex. These rocks contain albite, riebeckite, zircon, fluorite and Fe-Mn carbonate. Miller (1986) reported that the more mineralized 'exotic-rich' peralkaline granite from the Strange Lake Zr-Y-Nb-Be-REE deposit, has a high total REE content and is strongly enriched in HREE.

The Cross Hills Plutonic Suite has many geological, geochemical and metallogenic similarities to five posttectonic alkali to peralkaline granites from Saudi Arabia, which are of similar late Precambrian age. These granites are typically enriched in Nb, Ta, Sn, REE, Y, Th, U and Zr and in some cases contain economically important concentrations of some of these elements (Drysdall *et al.*, 1984; Jackson *et al.*, 1985). In particular, the Ghurayyah and Jabal Tawlah deposits show extreme enrichment in HREE and the Jabal Sa'id deposits are enriched in HREE and sporadically enriched in LREE (Figure 6) (Harris and Mariner, 1980; Drysdall *et al.*, 1984).

It has been suggested that under alkaline and oxidizing conditions HREE are partitioned into F^-- or CO_3^{2-}-bearing fluids (Kosterin, 1959; Flynn and Burnham, 1978; Kerrich and Fryer, 1979; Taylor *et al.*, 1981). Harris and Mariner (1980) demonstrated that fluorine is enriched in the Ghurayyah Granite and shows a positive correlation with HREE contents. They further suggested that HREE were maintained as fluoride complexes until the final stages of crystallization of the granite.

The peralkaline granite phase of the Cross Hills Plutonic Suite is actually depleted in fluorine compared to the biotite granite (Table 1). Although CO_2 contents of the granite are generally low (Table 1), the presence of an irregular 2- by 5-m vein of hematite-stained calcite within the peralkaline granite (centre of Figure 2), indicates that carbonate-rich fluids were formed during the hydrothermal evolution of the Cross Hills Plutonic Suite (Tuach, *in preparation)*. This suggests that CO_3^{2-} may have been more important in the concentration of elements in the Cross Hills Plutonic Suite as suggested by Taylor *et al.* (1981) for the Topsails complex.

The distribution of units, intrusive relationships, arrangement of textures and compositional variation within the Cross Hills Plutonic Suite, suggest a zoned magma

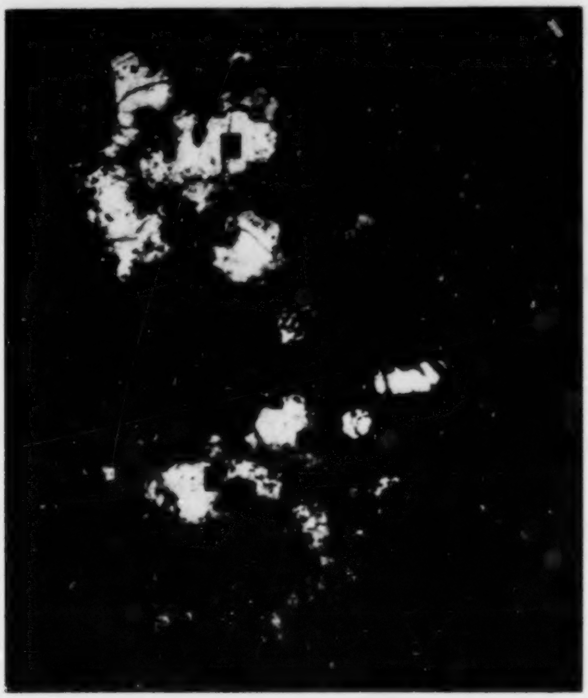

Plate 6. *SEM backscatter photograph showing botryoidal zircon (Z) thorite (TS) and thorium oxide (T); sample JT-067.*

chamber model (Figure 3), in which a dominantly liquid peralkaline magma was underlain by a similarly liquid (biotite) granite magma during the evolution of the pluton. Pressure-quenching resulted in a 'chill zone', which transected the compositional boundary between the two magmas (Figure 3). Crystallization of the peralkaline magma resulted in fluid-overpressure and transportation of Zr, Y, Nb, REE etc., to new depositional sites in the granite (Tuach, *in preparation*).

The inconsistent enrichment of Ce presumably is controlled by its oxidation state. Ronov *et al.* (1967) noted that tetravalent cerium has chemical properties that are more similar to those of the HREE than the other LREE. Its fractionation from the other LREE in some samples indicates that conditions were locally favourable to the maintenance of Ce in the +4 state. This is supported by the ubiquitous presence of hematite and its abundance in tuffisite, which indicates that oxidizing conditions prevailed during the final stages of granite formation.

CONCLUSIONS

The peralkaline granite phase of the Cross Hills Plutonic Suite is host to numerous small Zr-Nb-Y-REE mineral occurrences. The mineralogy of the occurrences and the distribution of mineralization in aplite dykes, tuffisite and fractures, indicate a magmatic-hydrothermal origin. The peralkaline granite provided a source for the mineralizing elements, which include Zr, Nb, Y, Sn, Th, Ce and heavy REE. Concentration and mobilization of these elements was

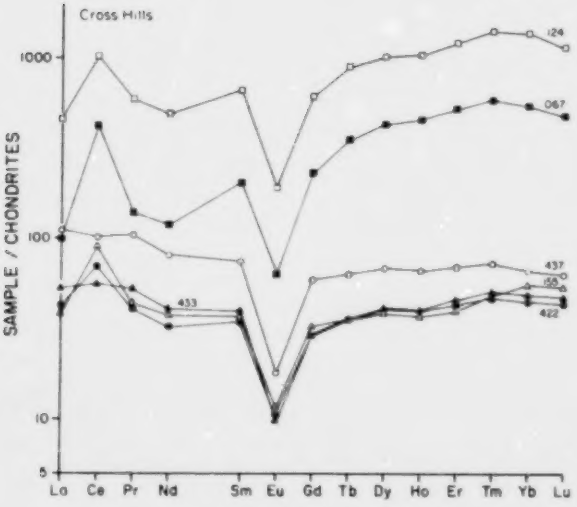

Figure 5. *Rare-earth element profiles of 4 peralkaline granite and 2 mineralized peralkaline granite samples; data are normalized to the chondrite values of Evensen et al. (1978).*

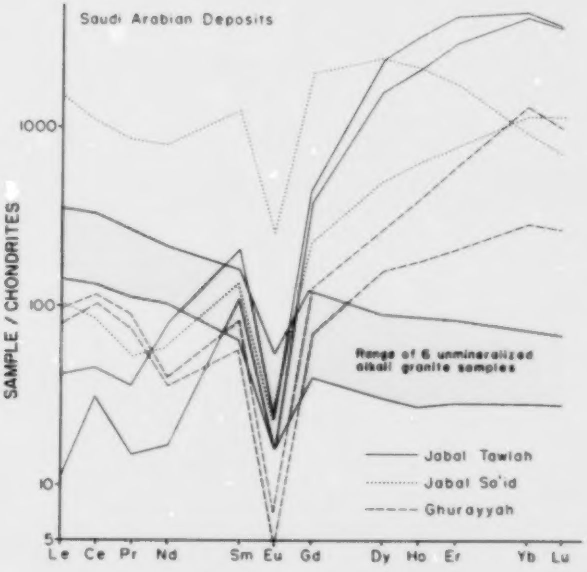

Figure 6. *Rare-earth element profiles of mineralized peralkaline granite samples from three Saudi Arabian deposits—Ghurayyah, Jabal Tawlah and Jaba Sa'id; also shown is the range for 6 unmineralized Saudi Arabian peralkaline granite samples; data from Drysdall et al. (1984).*

aided by the prevalence of oxidizing, alkaline and possibly carbonate-rich fluids.

The similarities between the geology, age and mineralization of the Cross Hills Plutonic Suite and that of the major deposits in Saudi Arabia suggest some potential for major Zr-Y-Nb-REE mineralization in the Cross Hills Plutonic Suite. The presence of similar granites in the Avalon Zone in Newfoundland and the remainder of the Appalachian Orogen points to other potential areas for such mineralization.

ACKNOWLEDGMENTS

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METALLOGENY OF THE CONNAIGRE BAY GROUP, SOUTHERN NEWFOUNDLAND

W.A. Sears[1] and C.F. O'Driscoll
Mineral Deposits Section

ABSTRACT

The Late Precambrian Connaigre Bay Group is host to several small, massive sulphide occurrences and showings. They are located within a limited stratigraphic horizon, consisting of the upper section of the Tickle Point Formation and the conformably overlying and predominantly mafic volcanic lower section, of the Sam Head Formation.

The major showing located at Winter Hill is situated in a carbonate–calc-silicate lens, which is in turn located within mafic volcanics. Sphalerite, galena and chalcopyrite are the main ore minerals whereas pyrite, sphalerite and pyrrhotite are the dominant sulphides. Epidote and chlorite alteration are widespread in the mafic volcanics around the mineralized zone, and within the mineralized lens sericite, calc-silicate and chlorite alteration assemblages dominate.

Most of the other minor occurrences appear in the upper Tickle Point Formation and generally consist of massive pyrite lenses containing only trace amounts of base-metal sulphides. An exception to this is at Frenchman Head, where base-metal sulphides locally make up 5 percent of the rock.

A volcanogenic–exhalative genesis is suggested by most of the macroscopic and microscopic observations. The lithological setting, alteration assemblages and ore mineral abundances, are a few of the criteria that conform to a volcanic origin.

INTRODUCTION

The Connaigre Bay Group lies in the Gaultois (1M/12) map area and is accessible by a paved road (Route 360) to Harbour Breton and an unpaved road (Route 364) to Hermitage.

The 1988 field season was devoted to a metallogenic study of the Connaigre Bay Group. Known and newly found base-metal occurrences were sampled and mapped in detail. Emphasis was placed on the more extensively worked Winter Hill and Frenchman Head occurrences (Figure 1).

The area was first mapped by Widmer (1950). Greene (1975) mapped the Harbour Breton map area, and Greene and O'Driscoll (1976) mapped the Gaultois map area, southeast of Hermitage Bay on a 1:50,000 scale. The area was also the subject of an M.Sc. thesis study by O'Driscoll (1977), the results of which were reported by O'Driscoll and Strong (1979).

The topography is quite rugged, having steep valley walls grading into undulating upland regions. Bedrock exposure is good in the upland areas except in places where bogs fill the topographic lows. The valley slopes are generally covered with a thin veneer of vegetation near the top and the lower sections are usually tree covered.

GENERAL GEOLOGY

The Hermitage Peninsula is located along the southwestern margin of the Avalon Zone (Williams, 1979). It is separated from the metamorphosed and polydeformed rocks of the Gander Zone by the Hermitage Bay Fault.

The oldest rocks on the peninsula are silicic–mafic volcanics and sedimentary rocks of the Late Precambrian Connaigre Bay Group (O'Driscoll and Strong, 1979). The Late Precambrian age is inferred from a lithological correlation with the Long Harbour Group to the east (Greene and O'Driscoll, 1976). All the rocks of the Connaigre Bay Group are separated from other sedimentary sequences by either faults or igneous intrusions (O'Driscoll, 1977).

The Connaigre Bay Group is intruded by the Hermitage Complex, which consists of diorite, gabbro and granite (O'Driscoll, 1977), and outcrops along the southwestern portion of the Hermitage Peninsula. Numerous felsic and mafic dykes associated with the Hermitage Complex cut and appear to feed the volcanics of the Connaigre Bay Group. The Straddling Granite is a small intrusion associated with the Hermitage Complex (O'Driscoll, 1977), and is located near the head of Hermitage Bay.

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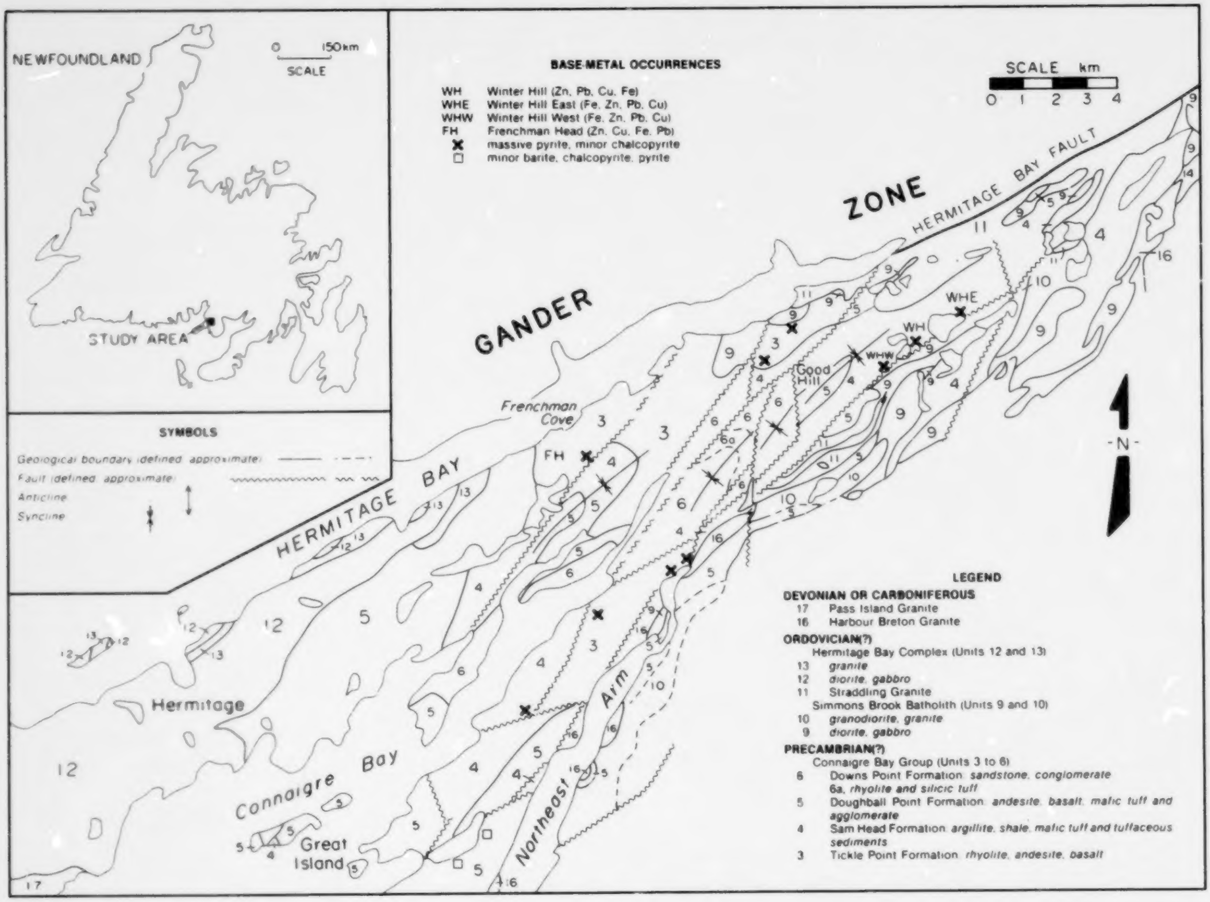

NEWFOUNDLAND
0 150km
SCALE
STUDY AREA
SYMBOLS
Geological boundary (defined, approximate)
Fault (defined, approximate)
Anticline
Syncline
BASE-METAL OCCURRENCES
WH Winter Hill (Zn, Pb, Cu, Fe)
WHE Winter Hill East (Fe, Zn, Pb, Cu)
WHW Winter Hill West (Fe, Zn, Pb, Cu)
FH Frenchman Head (Zn, Cu, Fe, Pb)
✖ massive pyrite, minor chalcopyrite
☐ minor barite, chalcopyrite, pyrite
SCALE km
0 1 2 3 4
GANDER
ZONE
HERMITAGE BAY FAULT
HERMITAGE BAY
Frenchman Cove
FH
Good Hill
WH
WHE
WHW
Hermitage
Connaigre Bay
Great Island
Northeast Arm
-N-
LEGEND
DEVONIAN OR CARBONIFEROUS
17 Pass Island Granite
16 Harbour Breton Granite
ORDOVICIAN(?)
Hermitage Bay Complex (Units 12 and 13)
13 granite
12 diorite, gabbro
11 Straddling Granite
Simmons Brook Batholith (Units 9 and 10)
10 granodiorite, granite
9 diorite, gabbro
PRECAMBRIAN(?)
Connaigre Bay Group (Units 3 to 6)
6 Downs Point Formation: sandstone, conglomerate
6a, rhyolite and silicic tuff
5 Doughball Point Formation: andesite, basalt, mafic tuff and agglomerate
4 Sam Head Formation: argillite, shale, mafic tuff and tuffaceous sediments
3 Tickle Point Formation: rhyolite, andesite, basalt

Figure 1. *Geology of the Hermitage Peninsula area (after Greene and O'Driscoll, 1976).*

The Hermitage Complex is intruded by the Pass Island Granite, which is similar to the Ackley Granite (Widmer, 1950) and considered to be about the same age (Devonian).

The Connaigre Bay Group

The Connaigre Bay Group is divided into four formations (O'Driscoll, 1977; Colman-Sadd *et al.*, 1979).

The basal formation is the Tickle Point Formation, which outcrops between the upper reaches of Connaigre Bay and Northeast Arm (Harbour Breton) and between Frenchman Cove and the head of Hermitage Bay (Figure 1). It consists of purple to pink, massive flow-banded and autobrecciated rhyolite and interbedded massive green andesite and basalt (O'Driscoll, 1977). The base of the Tickle Point Formation is not exposed, but, from outcrop exposures, the thickness of the formation is estimated to be at least 500 m.

The Sam Head Formation (Colman-Sadd *et al.*, 1979) conformably overlies the Tickle Point Formation. Close to Connaigre Bay, the Sam Head Formation consists of laminated green and grey argillite and has purple conglomerate, sandstone and shale at the base (O'Driscoll, 1977). Limestone lenses and limy argillite layers are rare. Southeast of upper Hermitage Bay it consists of interbedded mafic tuffs and tuffaceous sedimentary rocks (O'Driscoll, 1977). Carbonate–calc-silicate lenses are locally present and are important as a site for base-metal mineralization (see below). The total thickness of the Sam Head Formation is estimated to be at least 300 m (O'Driscoll, 1977).

The Doughball Point Formation conformably overlies the Sam Head Formation. It outcrops around Connaigre Bay and the Northeast Arm (Harbour Breton). It also appears as small conformable and fault-bounded lenses in the Good Hill area. The formation comprises green to grey massive andesite and basalt, fine to coarse, mafic tuff and agglomerate and minor interbedded silicic flows and tuffs. The Doughball Point Formation is the thickest formation in the Connaigre Bay Group (O'Driscoll, 1977) and is about 1500-m thick.

The Downs Point Formation is the uppermost sequence in the Connaigre Bay Group and conformably overlies the

Doughball Point Formation. This formation consists of red, purple and grey, graded and cross-bedded sandstone and pebble to cobble conglomerate. It also contains red, thinly laminated argillite and pink to purple massive rhyolite and silicic tuff (O'Driscoll, 1977). It outcrops along the northwest shore of upper Connaigre Bay and to the northeast of the head of Connaigre Bay. The estimated thickness of the Downs Point Formation is 1000 m, but faulting within the thickest exposed section (southwest of Good Hill) may have caused repetition or omission of some of the unit (O'Driscoll, 1977).

MINERALIZATION

Sam Head Formation

Field Relationships. The Sam Head Formation hosts the Winter Hill showings (Winter Hill (main showing), Winter Hill East and Winter Hill West) near a faulted contact with the Simmons Brook Batholith (Figure 1). The three showings occupy the same stratigraphic position along approximately 3 km of strike length. The Winter Hill showing occurs in a carbonate–calc-silicate lens, within mafic tuff, near the top of the Sam Head Formation (Figure 2). The Winter Hill East and Winter Hill West showings are in silicified argillite or fine grained tuffaceous rocks. All the showings are cut by felsic and/or mafic dykes or sills.

Sphalerite, pyrite, pyrrhotite, galena and chalcopyrite are the main sulphide minerals at the Winter Hill showing. There are also minor to trace amounts of magnetite. The mineralized zone, approximately 2- to 4-m thick, consists of disseminated to massive banded sulphides (Plate 1) underlain(?) by a zone of disseminated to massive pyrite and minor chalcopyrite, which has a maximum thickness of 4 to 5 m, but is irregular in shape. Bands of sphalerite (up to 10-cm thick) (Plate 2) occur discontinuously along the upper contact of the carbonate–calc-silicate unit. These sphalerite bands are overlain by either felsic or mafic intrusives or mafic tuffs. Magnetite also occurs along this upper contact, but only where the carbonate–calc-silicate lens is in contact with felsic intrusives.

The sphalerite bands grade downward into a calc-silicate facies containing thin laminae of sphalerite and increasing amounts of other sulphides. Underneath this is a discontinuous breccia unit (Plate 3) having fragments up to 0.5 m in diameter that are locally enveloped by thin bands of sphalerite. Gossanous material is common where sphalerite and other sulphides are present in the breccia. The gossan is generally rich in pyrite, chalcopyrite and sphalerite and has pervasive malachite and smithsonite staining. Carbonate having lesser calc-silicates underlie the breccia unit. Generally, the carbonate is unmineralized and exhibits contorted banding (Plate 4) (alternating carbonate- and sericite-rich layers), but near the breccia unit there is a thin (approximately 10 cm) bed of massive layered pyrite, sphalerite and galena. The attitude of this bed conforms with the local bedding.

Outcrop exposures at Winter Hill indicate that the carbonate–calc-silicate lens is at least 120 m in length, and between 5- and 10-m thick. Intersections in drill core indicate the unit, which dips between 30 and 40° to the northwest, is continuous, but is thinner about 150 m down dip.

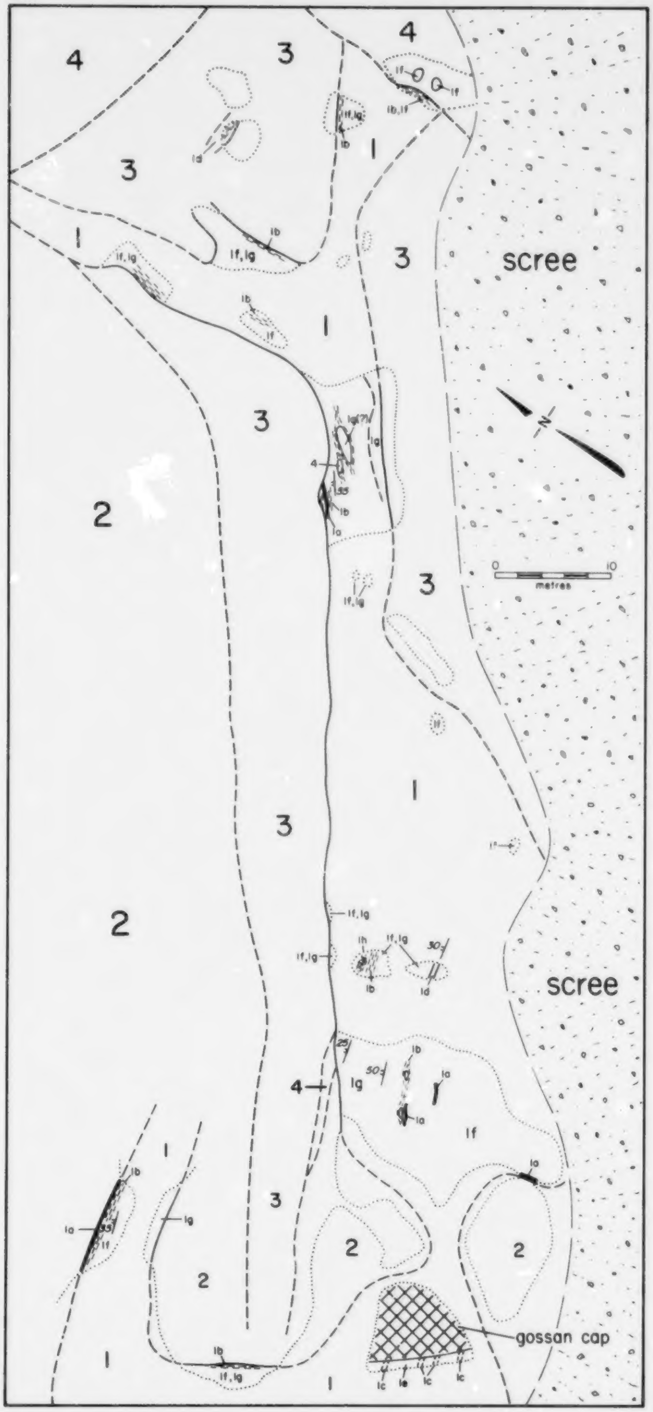
Winter Hill East and Winter Hill West are two small base-metal showings approximately 2 km northeast and 1.2 km southwest of Winter Hill, respectively. Both showings are hosted by silicified argillites or fine grained tuffaceous rocks. Mineralization at Winter Hill East consists of disseminated, semi-massive and massive pyrite and minor amounts of sphalerite and galena. The trench here is almost 30-m wide, but the best sulphide mineralization occurs over 5 to 10 m in the central portion of the trench. The host rock is bounded by a vesicular mafic flow or dyke along the western extension of the trench. The eastern section of the trench is mainly sheared argillite and disseminated pyrite.

The mineralized zone at Winter Hill West is smaller than the zone at Winter Hill East and consists almost entirely of semi-massive to massive pyrite. There is some banded pyrite and locally, disseminated sphalerite and galena are present.

Petrography. The mineralogy of the mineralized zones in the Sam Head Formation was studied from outcrop specimens and drill core samples. The local mafic lithic-crystal tuffs contain abundant fine grained plagioclase and lesser amounts of carbonate, pyroxene, and opaques as matrix components. Potassium feldspar, along with minor quartz, plagioclase, and hornblende and/or pyroxene make up the crystals, whereas the lithic clasts, generally, have the same mineralogy as the tuff.

Alteration is quite extensive within the tuffs. The matrix appears to be the least altered, but the fine grained crystals are cloudy and in places altered to assemblages of epidote, chlorite, and carbonate. Generally, the lithic clasts are more altered and contain similar assemblages. The K-feldspar crystal fragments are anhedral to subhedral and are commonly saussuritized. In a few cases, crystal overgrowths of a fibrous mineral obscures contacts between the crystal and the surrounding matrix. Plagioclase crystals show albite twinning and are also partly saussuritized. The hornblende and/or pyroxene crystals are altered to a combination of epidote and chlorite, whereas the quartz crystals are essentially fresh but show undulose extinction. Epidote and chlorite alteration is more prevalent closer to the ore zone, whereas sericite, carbonate and albite (?) alteration is more abundant farther away from the ore zone. This may be a direct result of the increase in mafic crystals as the ore zone is approached.

The carbonate–calc-silicate lens, which hosts the base-metal mineralization at Winter Hill (main showing) consists mainly of calcite and/or dolomite and lesser amounts of tremolite, diopside, sericite and quartz. Pyrite, sphalerite and rutile (hematite?) locally constitute between 1 and 2 percent



LEGEND

ORDOVICIAN OR OLDER

Hermitage Bay Complex

4 MAFIC DYKES/SILLS

3 FELSIC DYKES/SILLS

LATE PRECAMBRIAN

Sam Head Formation (Connaigre Bay Group)

2 MAFIC TUFF

1 CARBONATE/CALC-SILICATE LENS

- 1a semi-massive to massive sphalerite
- 1b thin sphalerite bands
- 1c semi-massive to massive pyrite (± cpy, po, gn, sph)
- 1d semi-massive to massive banded sulphides
- 1e semi-massive to disseminated pyrite
- 1f carbonate/marble
- 1g calc-silicate
- 1h breccia

Symbols

Geological boundary (known, approximate).. ⁄ ⁄

Major outcrops... ⬭

Sulphide/Carbonate banding.......................... 50

Figure 2. *Geology of the Winter Hill showing.*

Plate 1. *Thinly banded massive sulphides (subunit 1d).*

Plate 2. *Band of semi-massive to massive sphalerite (subunit 1a).*

of the rock. Near the lower contact with the mafic tuff, there is secondary serpentine, chlorite, epidote and possible talc. Relict olivine crystals, now mostly serpentine, are also present.

Tremolite and diopside are the major constituents of the calc-silicate facies associated with smaller percentages of quartz, carbonate and sericite. Epidote and chlorite are common alteration products and sericite, plagioclase and rutile are rare.

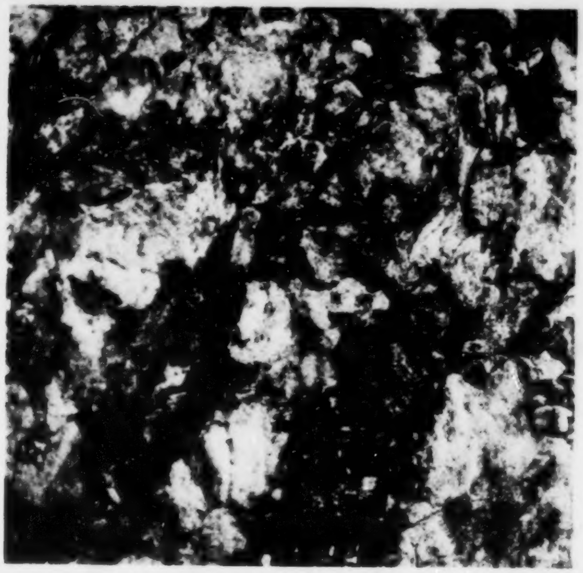


Plate 3. *Carbonate breccia unit at Winter Hill.*

Plate 4. *Contorted banding within the carbonate lens.*

Sphalerite and pyrite are the most abundant and widespread opaque minerals present in the Winter Hill showing. Chalcopyrite is usually present, associated with sphalerite and pyrite, but only in minor amounts. Galena, pyrrhotite and magnetite are also present in small quantities but are more localized. Digenite and chalcocite(?) occur as rare disseminated grains.

Sphalerite and pyrite occur in all mineralized areas, but sphalerite tends to be concentrated near the top of the carbonate–calc-silicate lens and pyrite is more common below the sphalerite-rich layers. Sphalerite, generally, occurs as medium- to coarse-grained anhedral crystals commonly containing chalcopyrite inclusions. Pyrite is usually present as subhedral to euhedral, fine- to medium-grained cubes, but rarely displays a colloform-like texture. Some of the pyrite cubes show evidence of fracturing along their boundaries whereas others are partially replaced by pyrrhotite(?). Sphalerite cores are also present within some of the pyrite crystals.

Chalcopyrite, present as inclusions in sphalerite, also occurs as fine grained crystal masses normally in close proximity to pyrite. Fine grained digenite and chalcocite, where present, are associated with chalcopyrite. Galena and pyrrhotite tend to be concentrated near the upper contact with the felsic unit. Galena occurs as anhedral, fine- to medium-grained crystals usually concentrated in thin (millimetre scale) layers. Pyrrhotite is present as small disseminated crystals and as a possible alteration product of pyrite.

Magnetite is almost solely limited to the contact zone between the carbonate–calc-silicate unit and the upper felsic unit. It occurs as thin discontinuous layers or as small crystal aggregates.

Tickle Point Formation

Field Relationships. Mineralization at Frenchman Head is contained in felsic volcanics of the lower Tickle Point Formation. The showings contain only minor disseminated and banded base-metal sulphides (up to 5 to 10 percent). The silicified host rocks in the mineralized area commonly contain quartz 'eyes' in a matrix of fine grained quartz, sericite and chlorite. Pyrite, sphalerite, galena and chalcopyrite, generally, occur as disseminated grains, but are rarely concentrated in thin bands. Pyrite is the dominant sulphide in these thin bands. Chalcopyrite, pyrite and galena ± chalcocite also occur along some fractures. The mineralized zone is commonly overlain by carbonatized and chloritized mafic volcanics.

On the surface, the sulphide-rich zones are exposed in three lateral mineralized outcrops that apparently occupy the same stratigraphic level; these zones are between 5- and 7-m thick and probably up to 400 m long. The sulphide mineralization is cut by porphyritic mafic dykes.

Massive and massive-banded pyrite occur at several other localities in the Tickle Point Formation. These occurrences are commonly barren of base-metal sulphides but chalcopyrite is present as a minor to trace constituent in some localities. The massive pyrite lenses are up to 2- to 3-m thick and may be continuous for 500 to 1000 m along strike. They are hosted by fine grained felsic volcanic rocks, felsic tuffs, silicified argillites and/or cherts. Minor amounts of mafic tuff are not uncommon at most occurrences.

Petrography. Sulphide-rich zones in the Tickle Point Formation are contained essentially within fine grained felsic rocks. These rocks are characterized predominantly by fine grained quartz. Aggregates of quartz crystals commonly form small spheres and may indicate a spherulitic texture. Plagioclase is present as phenocrysts and in the matrix. Fresh plagioclase normally displays albite twins, whereas some show oscillatory zoning (possible sanidine). However, the majority of the plagioclase is altered to either chlorite, sericite, or epidote. Potassium feldspar is also present but may be difficult to distinguish from untwinned and singularly twinned plagioclase. Pyroxene and hornblende occur rarely and are usually partly or wholly altered to epidote or chlorite.

At Frenchman Head, pyrite comprises approximately 80 percent of the total opaque mineralogy; sphalerite and chalcopyrite constitute the remainder. The pyrite is present as fine- to medium-grained disseminated crystals. Sphalerite grains are usually larger than the pyrite grains and commonly contain inclusions of chalcopyrite. Chalcopyrite also occurs as small discrete crystals.

Away from the Frenchman Head showing, pyrite is essentially the only opaque mineral. Some areas locally contain 1 to 2 percent sphalerite and chalcopyrite combined. The pyrite occurs as semi-massive to massive-layered, fine grained subhedral crystals. Sphalerite and chalcopyrite are generally anhedral and smaller in size. Hematite is present as coatings around pyrite crystals or as thin veinlets (possibly oxidized pyrite veins), in places containing rutile.

Doughball Point Formation

Field Relationships. In the Doughball Point Formation there is a minor occurrence of chalcopyrite and barite. Chalcopyrite is present as isolated patches in and around some mafic dykes whereas both barite and chalcopyrite are present in carbonate and quartz–carbonate veins cutting the mafic tuffs. Barite is limited mainly to the calcite–quartz veins, but only two veins in two different localities contain barite. Semi-massive pods of hematite are also located in some veins.

Petrography. The mafic dykes are composed essentially of equigranular plagioclase and pyroxene. Some of the pyroxenes are altered slightly to chlorite and epidote. The larger plagioclase crystals are rimmed by small pyroxene crystals. Potassium feldspar is also a minor component. One mineralized vein contains garnet and pyroxene.

The intruded mafic tuffs have been altered to a fine grained assemblage of calcite, epidote and other calc-silicate minerals.

DISCUSSION

Massive sulphide showings and occurrences within the Connaigre Bay Group are concentrated near the top of the

Tickle Point Formation and at the base of the conformably overlying Sam Head Formation. Macroscopic and microscopic examinations of these mineralized zones suggest a volcanic–exhalative origin. The criteria leading to this conclusion are as follows:

- a close association with volcanic and volcaniclastic rocks, fine grained sedimentary rocks and chemical precipitates (i.e., carbonates and cherts);
- the sulphide lenses are restricted to a limited stratigraphic package and were probably formed over a short interval of time;
- the common occurrence of thin sequences of massive pyrite, with or without base-metal sulphides;
- sulphide mineralogy dominated by pyrite and lesser amounts of sphalerite, pyrrhotite, galena and chalcopyrite;
- sphalerite and galena are the dominant ore sulphides in the upper section of the carbonate–calc-silicate lens at Winter Hill, whereas pyrite and minor chalcopyrite are more abundant in the lower section of the lens;
- local sericite and chlorite alteration; and
- granular texture displayed by much of the sulphides and some pyrite grains displaying rare colloform texture.

All the above criteria have not been found together, at any one occurrence or showing, but are a culmination of observations made during the study. At many localities, younger events (i.e., formation of skarnoid(?) minerals within the carbonate–calc-silicate lens at Winter Hill) have obscured original textures and also, possibly, have altered the primary mineral assemblages.

ACKNOWLEDGMENTS

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VOLCANIC ROCK GEOCHEMISTRY AS A GUIDE FOR MASSIVE SULPHIDE EXPLORATION IN CENTRAL NEWFOUNDLAND

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Mineral Deposits Section

ABSTRACT

Preliminary results from whole rock geochemical studies in the Wild Bight, Lushs Bight, Pacquet Harbour and Victoria Lake groups and the Betts Cove (ophiolite) Complex, central Newfoundland, suggest that volcanogenic massive sulphide (VMS) deposits in these sequences are hosted by volcanic rocks of island-arc geochemical signatures. There is a widespread specific association of VMS deposits with strongly incompatible element-depleted volcanic rocks, interpreted as the products of hydrous remelting of refractory mantle sources. The ore-bearing sequences of island-arc derivation are generally stratigraphically overlain by, or structurally juxtaposed with, volcanic rocks of non-arc signatures.

The geochemical and stratigraphic evidence suggests that massive sulphide deposition took place during arc-rifting events. The rocks with island-arc signatures represent the rifting stage (and in some cases, part of the pre-rifting history of the arc). Rocks with non-arc signatures represent volcanic activity in the newly-established back-arc basins.

The specific association of massive sulphides with strongly depleted volcanic rocks suggests a genetic link. It is postulated that in the rifting environment, these very hot magmas rise quickly to shallow depths with much of their heat of fusion intact. This anomalous heat would fuel more vigorous and larger hydrothermal cells, resulting in the leaching of increased quantities of metals in the subsurface. The deep fractures associated with the rifting would focus the fluids on the rising side of the cells, allowing more efficient delivery of metals fluids to the ore depositional environment. The presence of these strongly depleted volcanic rocks in ancient island arcs is probably a significant indicator of the presence of massive sulphides, and may prove to be a useful tool in regional exploration.

INTRODUCTION

Recent improvements in trace-element analytical techniques have provided geochemists and petrologists with a greater range of geochemical and isotopic data for studying ancient igneous rocks. Using Instrumental Neutron Activation Analysis (INAA), Inductively Coupled Plasma Mass Spectrometry (ICP-MS), and improvements in X-Ray Fluorescence (XRF) techniques, rapid, precise determination of the High Field Strength Elements (HFSE) (Ta, Nb, Zr, Hf, Y), Low Field Strength Elements (LFSE) (Rb, Sr, Ba, Cs,U, Th) and the Rare Earth Elements (REE) are now possible at very low concentrations. Utilizing this capability, detailed geochemical studies of volcanic rocks have been initiated in the massive suphide districts of central Newfoundland. The aims of these studies are, a) to determine the nature of the paleotectonic environments that are represented by these volcanic rocks, and b) to investigate the relationships between volcanic rock geochemistry and the presence or absence of volcanogenic massive sulphide (VMS) deposits.

The approach applied here, is based on the hypothesis that volcanogenic sulphide mineralization in central Newfoundland may have taken place in response to a particular set of paleotectonic conditions, as is the case in well studied modern examples such as in the Kuroko District, Japan, or in Fiji (Sillitoe, 1982; Cathles *et al.*, 1983). If the favourable paleotectonic environment can be identified and characterized by the types of volcanic rocks that erupted while it was extant (again, based on comparisons with modern environments), then perhaps those characteristics could be used to focus and direct exploration in complex oceanic volcanic terranes.

In 1984, a pilot study of the relationships between volcanic rock geochemistry and volcanogenic sulphide deposits was begun in the Wild Bight Group, central Newfoundland (Figure 1), to determine whether volcanic rocks associated with volcanogenic sulphides showed any characteristic geochemical signatures that could be used as a guide for exploration (Swinden, 1984, 1985). The initial

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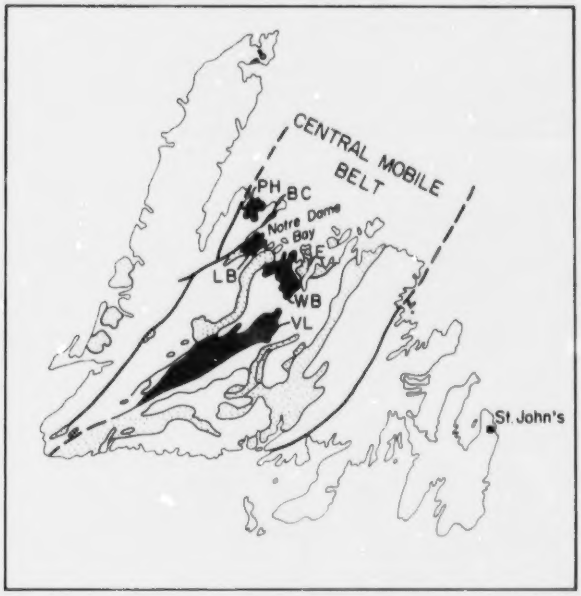

Figure 1. *Pre-Silurian vestiges of Iapetus Ocean in central Newfoundland (stippled). Sequences studied in this report are shown in black. WB–Wild Bight Group; LB–Lushs Bight Group; BC–Betts Cove Complex and Snooks Arm Group; PH–Pacquet Harbour Group; VL–Victoria Lake Group.*

success of this study (Swinden, 1985, 1987) led to the broadening of the scope of the project, to include other massive sulphide-bearing volcanic sequences in central Newfoundland. The objective of this comprehensive program is to identify and characterize consistencies in paleotectonic environments and geochemical signatures of the ore-associated rocks, for use as a tool in predictive metallogeny regionally across the Central Mobile Belt. Some of the preliminary results of this study are presented here.

Data from five sequences in north–central Newfoundland, the Wild Bight and Lushs Bight groups in Notre Dame Bay, the Betts Cove (ophiolite) Complex and the Pacquet Harbour Group on the Baie Verte Peninsula, and the Victoria Lake Group in south–central Newfoundland (Figure 1), suggest that massive sulphide mineralization occurred in these sequences during island-arc rifting. Host rocks in all sequences have a distinct geochemical signature of island-arc volcanism and are generally overlain by volcanic rocks generated from non-arc mantle sources. Within the rifted arc sequences, there is a close spatial association between VMS deposits and volcanic rocks that are anomalously depleted in incompatible elements, and apparently represent second-stage melting of refractory mantle sources. It is possible that this consistent association is an important indicator of tectonic events that trigger VMS deposition in these environments, and hence may have considerable application as an exploration tool.

METHODS

In this study, reliance is placed on elements that are likely to have been immobile through greenschist and lower amphibolite facies metamorphism at low water/rock ratios. These include the major elements TiO_2 and P_2O_5, the HFSE (Zr, Nb, Y, Ta, Hf) and the REE. Most LFSE (e.g., U, Rb, Sr, Cs) are too mobile to be useful in altered rocks. However, Th, although by definition a LFSE, is generally considered to be immobile under low-grade metamorphism.

In the following geochemical discussion, reference is made to 'arc signatures' and 'non-arc signatures'. Following Wood *et al.* (1979), Sun (1980), and many others, we consider oceanic volcanic rocks that result from magmatism influenced by subduction, to generally exhibit a marked depletion in the highly incompatible HFSE (Nb and/or Ta), and a marked enrichment in the LFSE elements (in altered rocks represented by Th) with respect to La. This is an empirical signature that is widely recognized in modern arc environments as characterizing arc magmatism, and although its origins remain somewhat contentious, the signature is clearly a magmatic phenomenon inherited from the source regions of the melts. The inter-element relationships that characterize this signature are generally considered to reflect mass transport from subducted sediment and/or altered oceanic crust from the subducting slab into magma source regions, in the overlying mantle wedge. They are manifested graphically on primitive mantle-normalized extended REE plots as a negative anomaly in Nb and a positive anomaly in Th with respect to La and Ce (Figure 2). Such an extended REE pattern constitutes an 'arc signature' in our terminology. Non-arc mantle sources will produce magmas having similar normalized abundances of La and Nb and normalized Th abundances that are similar to or less than La and Nb. Such a smooth pattern between Th and Ce constitutes a 'non-arc signature'.

To illustrate these relationships, a simple diagram has been devised, in which the ratio Nb/Th is plotted against Y (e.g., Figures 5, 8, 11, 14, 17, 19). The Nb/Th ratio is a measure of the extent of Nb depletion and Th enrichment in the magma source (i.e., it separates arc signatures with a low Nb/Th ratio from from non-arc signatures with higher Nb/Th ratio). Y serves as an indicator of the relative incompatible element depletion in the magma; very low Y contents (i.e., less than about 16 ppm) indicate highly depleted rocks that may be second stage melts. The stippled field of overlap between fields, results from uncertainties in analytical techniques and the gradational nature of the geochemical boundaries between these different rock types.

RESULTS

The Wild Bight Group

The Wild Bight Group (Figure 3) comprises a more than 8-km-thick sequence of Llanvirnian–Landeilian (Early

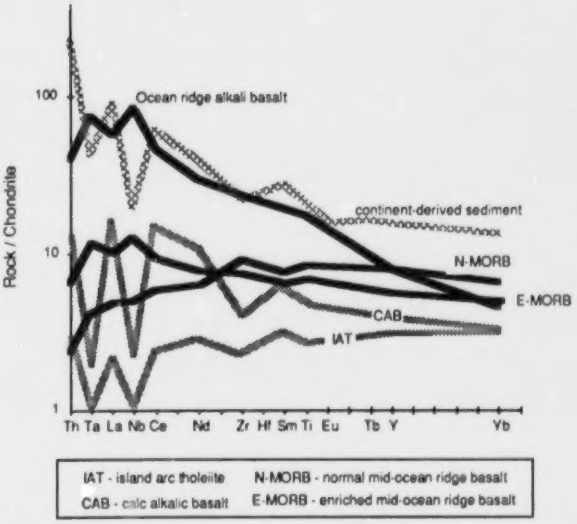

Figure 2. *Typical extended REE patterns for oceanic basalts of different types after Sun (1980), and a typical continent-derived oceanic sediment, after Hole et al. (1984). Island-arc rocks are stippled lines, non-arc rocks are black lines and sedimentary rocks are the hatched line. Note prominent negative Ta and Nb anomalies and high Th in the arc rocks and in the sediment.*

Ordovician) epiclastic, mafic and lesser felsic volcanic rocks and subvolcanic sills and dykes. Volcanic rocks, although comprising less than 25 percent of the stratigraphic record, occur at virtually all stratigraphic levels within the sequence in eleven separate volcanic accumulations. Mafic volcanic rocks comprise mainly pillow lava and lesser massive basalt flows, pillow breccia and fine grained hyaloclastite. Felsic volcanic rocks range from aphanitic rhyolite flows to crystal- and crystal-lithic tuff, and these form small, laterally restricted bodies within the dominantly mafic volcanic sequences.

The Wild Bight Group hosts four volcanogenic sulphide deposits (Figure 3). None have achieved production although one, the Point Leamington deposit, contains more than 13 mt of relatively low grade Cu–Zn-bearing massive sulphide (Walker and Collins, 1988). The Point Leamington and the Lockport deposits consist of massive sulphides capping extensive alteration stockworks. The former is hosted by felsic volcanic rocks, but has mafic pillow lavas approximately 200 m below in the footwall sequence; the latter is hosted by mafic volcanic rocks but is less than 300 m along strike from a small rhyolite body. The other two deposits, the Indian Cove and Long Pond prospects, comprise stockworks in felsic flow and pyroclastic rocks with no massive sulphide cap. The former has associated mafic volcanic rocks nearby in the stratigraphic sequence.

In a detailed petrological, geochemical and isotopic study of volcanic rocks in the Wild Bight Group, Swinden (1987) demonstrated a progressive change in the nature of the volcanic activity. The earliest volcanic rocks are dominantly LREE-enriched island-arc tholeiites (Figure 4). Later in the history of the Wild Bight Group, a new phase of volcanic activity began, which included continued eruption of island-arc tholeiites, but was characterised by the eruption of felsic volcanics of trondhjemitic composition and a suite of extremely incompatible element-depleted tholeiitic rocks ('strongly depleted arc', Figure 4). All of the volcanic rocks erupted to this point, carry a clear arc signature. Following eruption of these rocks, new mantle sources without the arc signature were apparently involved in the magmatism, and volcanic rocks at the top of the group carry a dominantly non-arc signature. This geochemical progression was interpreted (Swinden, 1987) to reflect a paleotectonic progression from an early stage of island-arc volcanism (the lowermost arc tholeiites), through an episode of arc-rifting (during which, the felsic volcanics and strongly depleted thoeliites were erupted), and culminating with the establishment of a back-arc basin (during which, the magmas with non-arc sources were erupted).

Volcanogenic sulphide deposits occur only in the central stratigraphic part of the Wild Bight Group and are, therefore, an integral part of the arc-rifting event. Furthermore, all are associated with strongly depleted tholeiites, the chemical characteristics of which are illustrated in Figure 4. These rocks contain very low concentrations of all incompatible elements (including less than 10 percent total REE) and are interpreted to have formed through complex, multi-stage hydrous remelting of the refractory mantle sources left after the removal of a first stage arc tholeiite melt (Swinden, 1987). All plot in Field 'C' of Figure 5. Wherever these rocks are found, volcanogenic sulphides are also found and conversely, no mineralization of this type is known to occur in the Wild Bight Group in their absence. Although it is true that volcanogenic sulphides are also closely associated with felsic volcanic rocks, the converse is not true. There are felsic volcanic bodies within the group that are apparently barren.

The Lushs Bight Group

The Lushs Bight Group (Figure 6) comprises a structurally complex sequence dominated by basaltic pillow lava having lesser pillow breccia and mafic tuffaceous rocks, and sheeted diabase dykes. Although generally considered to be part of a disrupted ophiolite (e.g., Dean, 1978; Kean, 1984, 1988), the Lushs Bight Group contains no cumulate plutonic elements of the ophiolite stratigraphy and ultramafic rocks occur only as very small fault-bounded fragments. The mafic volcanic rocks from different structural blocks in the Lushs Bight Group are generally homogeneous, both geologically and petrographically, and cannot be subdivided or correlated with confidence in the field. Felsic volcanic rocks occur locally in the western part of the group (Kean, 1988).

The Lushs Bight Group is overlain conformably by the Western Arm Group, a Lower Ordovician (MacLean, 1947) sequence of mafic pillow lava having lesser interbedded pyroclastic and epiclastic rocks. It has, historically, been interpreted as an island-arc sequence, built upon the Lushs Bight crust (Dean, 1978).

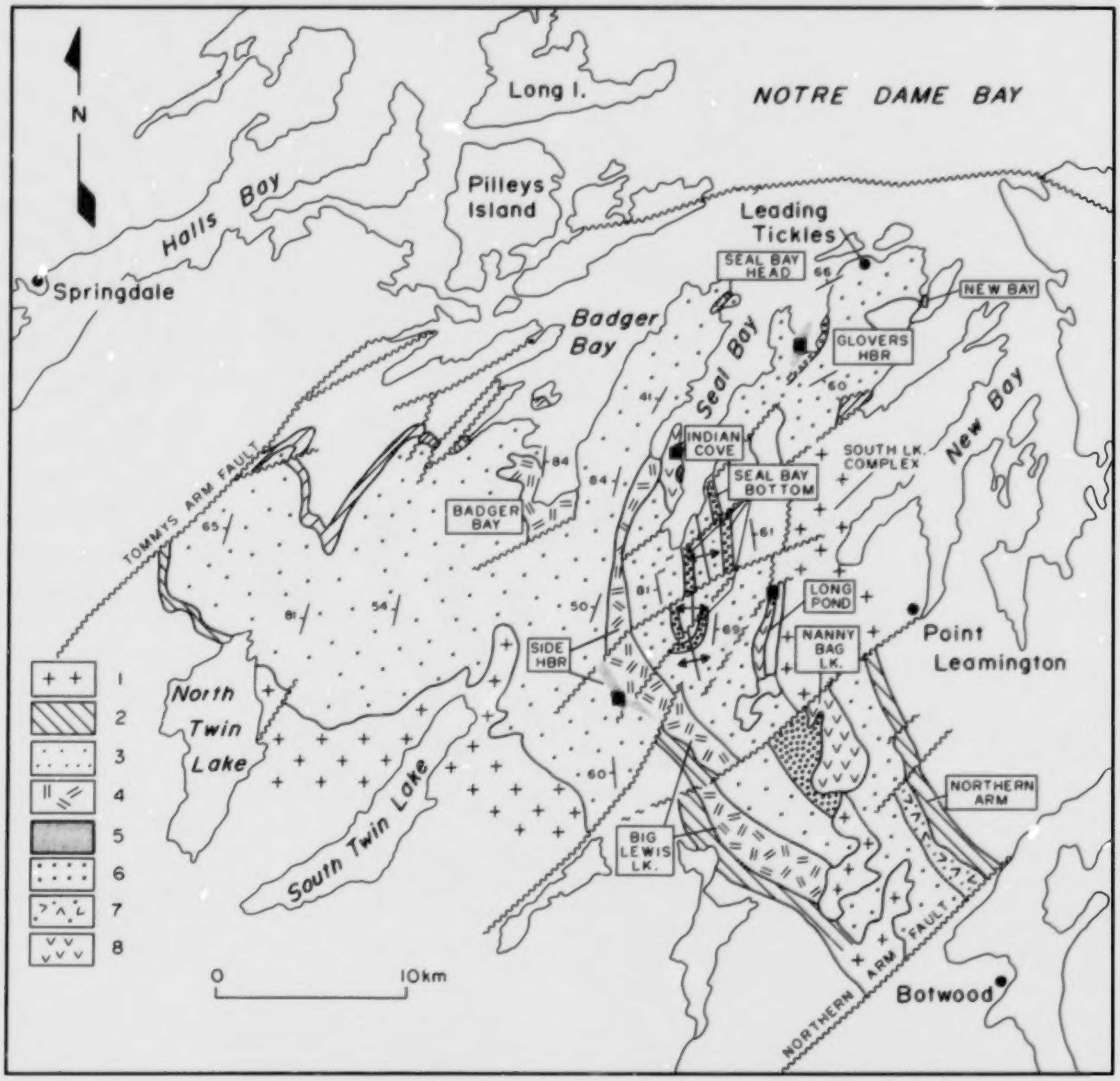
N
Long I.
NOTRE DAME BAY
Pilleys Island
Halls Bay
Springdale
Leading Tickles
SEAL BAY HEAD
GLOVERS HBR
NEW BAY
Badger Bay
Seal Bay
New Bay
INDIAN COVE
SEAL BAY BOTTOM
SOUTH LK. COMPLEX
BADGER BAY
TOMMYS ARM FAULT
LONG POND
NANNY BAG LK.
SIDE HBR
Point Leamington
North Twin Lake
South Twin Lake
NORTHERN ARM
BIG LEWIS LK.
NORTHERN ARM FAULT
Botwood
0
10km
1
2
3
4
5
6
7
8

Figure 3. *Geological map of the Wild Bight Group after Swinden (1987) showing the geochemical affiliation of the volcanic rocks. 1–post-Ordovician plutonic rocks; 2–Caradocian shale (Shoal Arm Formation); 3–epiclastic rocks; 4–non-arc volcanic rock; 5–strongly depleted arc-tholeiites; 6–LREE-enriched and LREE-depleted arc-tholeiites; 7–mixed arc/non-arc sequences; 8–rhyolites. Black squares are volcanogenic sulphide deposits.*

The Lushs Bight Group contains more VMS occurrences per square kilometre than any other group of rocks in central Newfoundland (Dean, 1978). Three of these, the Whalesback, Little Bay and Little Deer, have seen production within the past 20 years (Whalesback and Little Bay are in the 2.5 to 4 mt range). Several other deposits produced lesser amounts during the 'copper boom' of the late 1800's. A complete summary of the major VMS occurrences is found in Kean and Evans (1988a) and Swinden *et al.* (1988). All of the deposits consist dominantly of pyrite, chalcopyrite and sphalerite and commonly occupy chloritic shear zones. Stratiform massive sulphides apparently occur at more than one stratigraphic level within the volcanic sequence, and stockwork-like alteration can, locally, be seen in the volcanic rocks and in the underlying sheeted dyke sequence (Kean, 1984; Kean and Evans,1988a).

The basal unit of the overlying Western Arm Group comprises siliceous, chocolate-brown argillite and interbedded ferruginous chert, argillite and green sandstone.

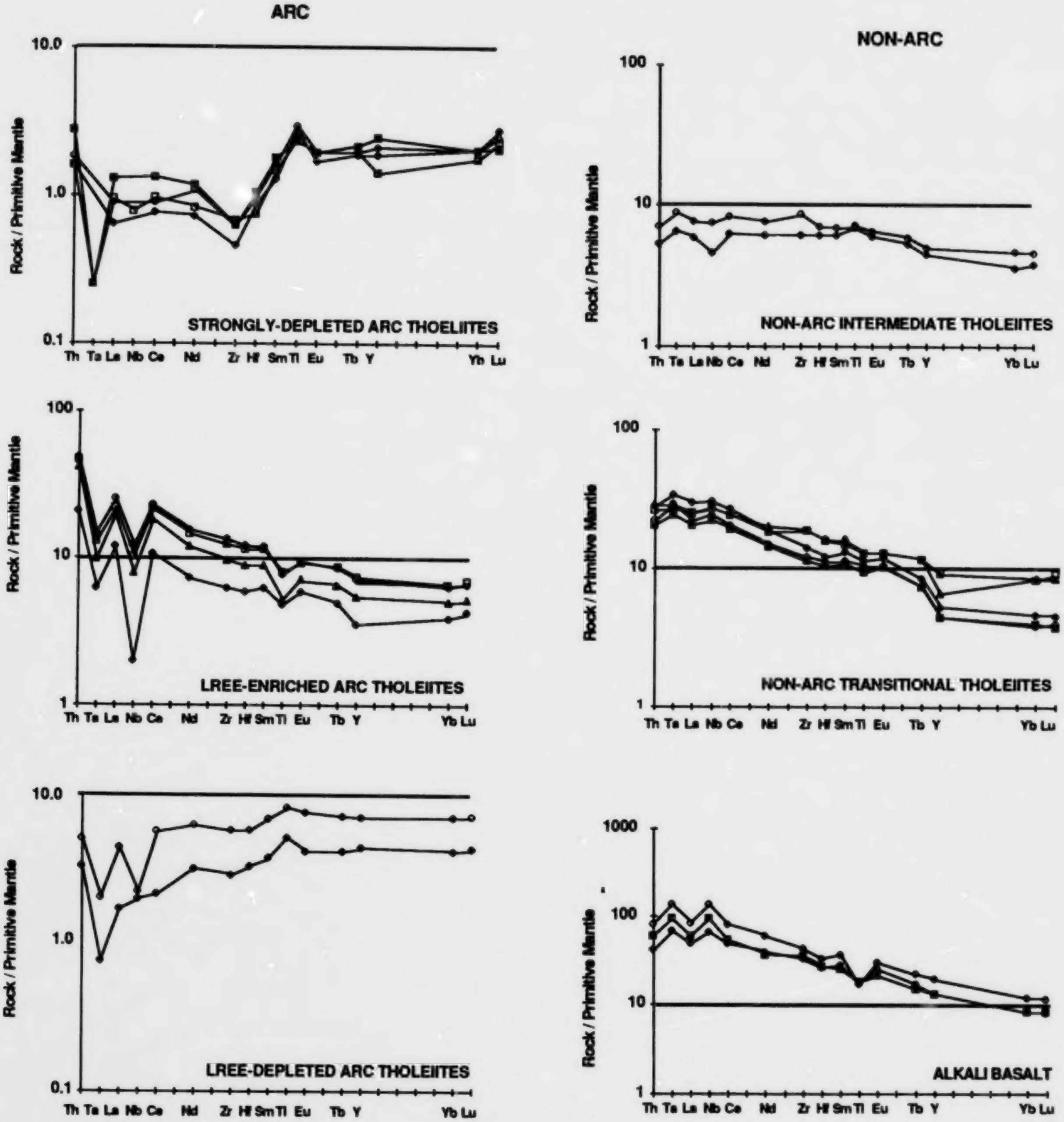

Figure 4. *Extended REE plots for volcanic rocks in the Wild Bight Group: Solid line on this and all subsequent extended REE plots is drawn at 10 x primitive mantle for ease of reference among plots. Primitive mantle normalizing values are: Th−0.088; La−0.63; Nb−0.65; Ce−1.59; Nd−1.21; Zr−9.8; Sm−0.399; Ti−1134; Eu−0.15; Gd−0.51; Y−3.9; Dy−0.75; Er−0.42; Yb−0.432; Lu−0.66.*

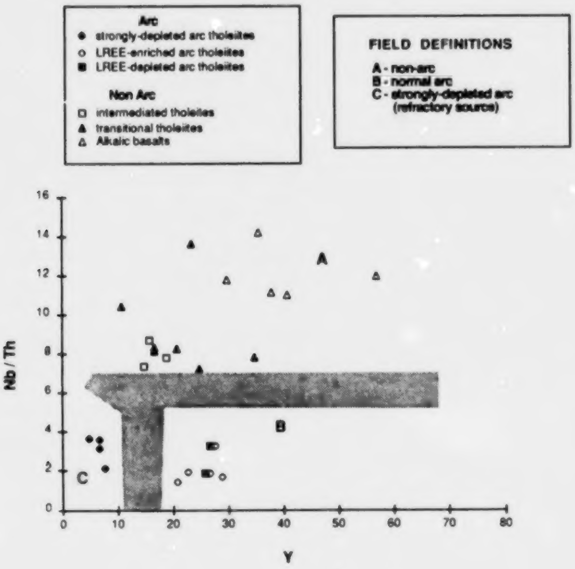

Figure 5. *Nb/Th versus Y for Wild Bight Group rocks: symbols are keyed to subdivisions illustrated in Figure 4. Stippled area is field of overlap.*

These sediments host narrow stratiform bands of massive pyrite, pyrrhotite, magnetite and rare chalcopyrite (Kean, 1984; Kean and Evans, 1988a). The bands ('Fox Neck-type' deposits of Swinden *et al.*, 1988), are essentially sulphide facies iron formations containing only minor base metals and no precious metals and are the only volcanogenic sulphides in the Western Arm Group.

Although the Lushs Bight Group has historically been interpreted as representing oceanic crust (Smitheringale, 1972) and the deposits as having formed at an oceanic or back-arc spreading centre (Kean, 1984), recent geochemical studies suggest that the story is much more complex (Jenner *et al.*, 1988). Several basalt units can be recognized in the volcanic rocks and in the dykes on the basis of their geochemistry, and most have a clear arc signature (Figure 7). Types I-N, I-S and III comprise dominantly normal arc-tholeiites (although Type I-N may also include some non-arc rocks, see Figure 8). Type II basalts are probably also arc-tholeiites, but are more depleted in incompatible elements than Types I and III and plot in the field of overlap between normal and strongly depleted rocks on Figure 8. Types IV and V are geochemically identical, very depleted boninites having high MgO contents and concave-upward REE patterns (Figure 7). Based on the very low contents of incompatible elements, rocks from Types IV and V and probably also those of Unit II are interpreted as the products of hydrous remelting of refractory mantle sources.

The island-arc geochemical signatures in the associated tholeiites suggest that the volcanism and mineral deposition did not occur at a well-established spreading centre as previously thought, but more likely in a supra-subduction setting. In the overlying Western Arm Group, the basalts in the basal Skeleton Pond tuff comprise a mixed arc–non-arc sequence and pass upward into pillow lavas of the Big Hill basalt, which have a non-arc signature (Figures 7 and 8) (Jenner and Szybinski, 1987). The Lushs Bight and Western Arm groups may, therefore, record a rifted-arc to back-arc transition similar to that interpreted for the Wild Bight Group with most, if not all, of the Lushs Bight Group having formed during the arc-rifting stage.

VMS deposits in the Lushs Bight Group show a distinct affinity for the incompatible element-depleted Type II tholeiites and Type IV and V boninites (Figure 6). However, it is not clear whether all deposits are actually hosted by the depleted rocks. Several major deposits including Rendall–Jackman, Lady Pond and Whalesback, apparently occur in normal arc-tholeiites, although more detailed sampling is still needed in the immediate vicinity of the deposits to confirm this. It may be significant that the Lady Pond deposit is spatially associated with boninitic dykes. Any genetic model that attempts to relate the occurrence of VMS deposits to the presence of depleted magmatic rocks must take this into account.

The Betts Cove Complex

The Betts Cove Complex (Figure 9) is a complete ophiolite, which outcrops in an arcuate belt on the eastern side of the Baie Verte Peninsula. The east-facing stratigraphic sequence consists of basal ultramafic rocks that pass upward into mafic cumulate and plutonic rocks, sheeted diabase dykes and an upper unit of mafic pillow lavas (Upadhyay, 1973). The ophiolite is stratigraphically overlain by the Snooks Arm Group, a sequence of epiclastic and mafic volcanic rocks that Jenner and Fryer (1980) have interpreted to represent an oceanic-island environment.

Sun and Nesbitt (1978) and Coish and Church (1979) first recognized unusual Mg-rich, incompatible element-depleted lavas in the Betts Cove ophiolite. Coish *et al.* (1982) presented geochemical and Nd-isotopic data showing that these very depleted melts are associated with normal tholeiites in the stratigraphy, and presented Nd isotopic data in support of an origin in a supra-subduction setting. The data (Figure 10) show that these very depleted rocks are geochemically similar to boninites and, as suggested by Coish *et al.* (1982), must represent remelting of refractory sources.

Elsewhere in the Betts Cove sequence, tholeiitic rocks with REE patterns ranging from flat to LREE-depleted, and distinct negative Nb anomalies, are present. The geochemical patterns and Nd isotopic data presented by Coish *et al.* (1982) suggest that these probably represent the melting of normal mantle sources in an island-arc environment. The most strongly LREE-depleted rocks are similar to MORB, both geochemically and isotopically, and plot in the non-arc field in Figure 11. This suggests that mantle sources, which were not influenced by the subducting slab, were involved in their petrogenesis. In the stratigraphically overlying Snooks Arm

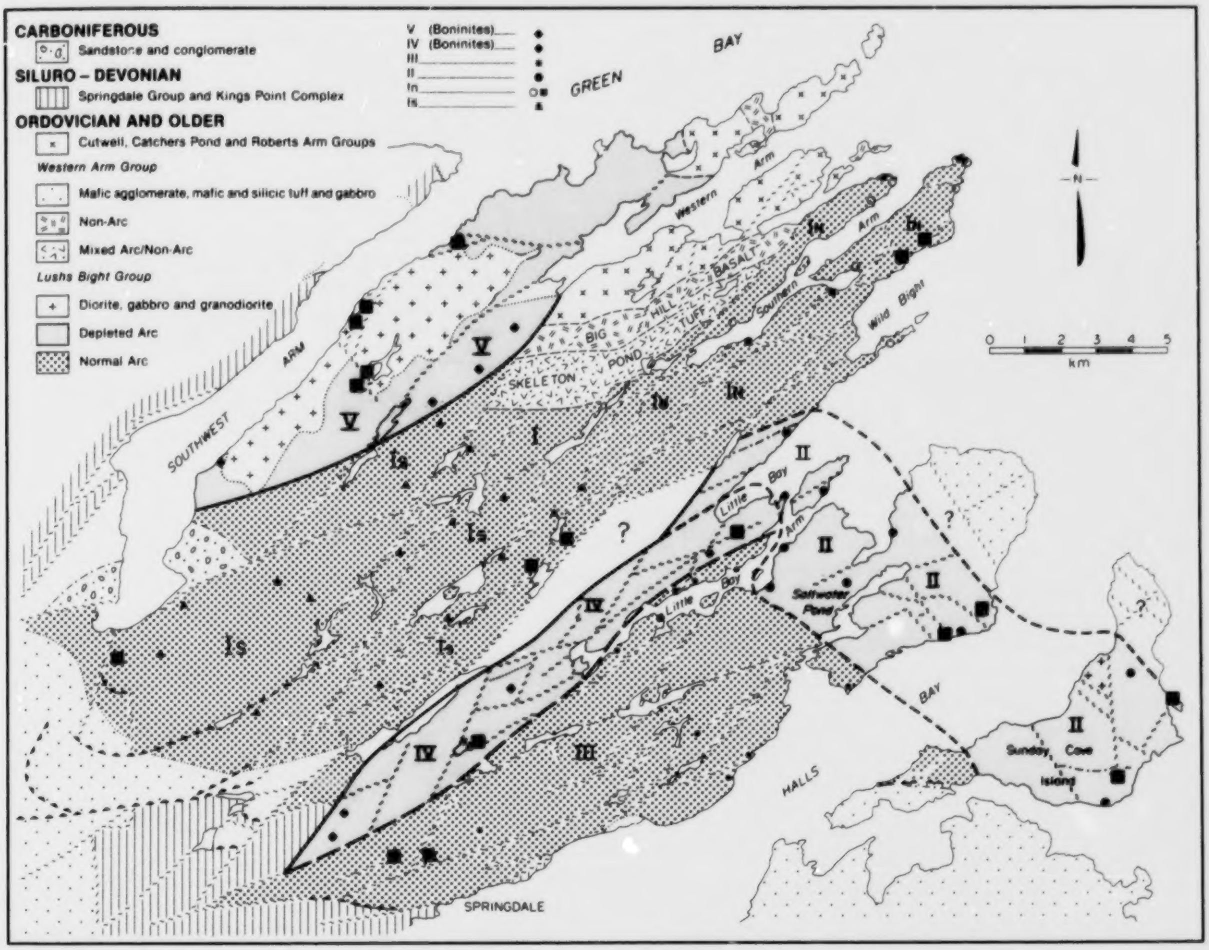
CARBONIFEROUS
Sandstone and conglomerate
SILURO – DEVONIAN
Springdale Group and Kings Point Complex
ORDOVICIAN AND OLDER
Cutwell, Catchers Pond and Roberts Arm Groups
Western Arm Group
Mafic agglomerate, mafic and silicic tuff and gabbro
Non-Arc
Mixed Arc/Non-Arc
Lushs Bight Group
Diorite, gabbro and granodiorite
Depleted Arc
Normal Arc
V (Boninites)
IV (Boninites)
III
II
In
Is
GREEN
BAY
Western
Arm
Southern
Arm
Wild Bight
BIG HILL BASALT
BIG POND TUFF
SKELETON
SOUTHWEST
ARM
Little Bay
Arm
Little Bay
Saltwater Pond
BAY
HALLS
Sunday Cove Island
SPRINGDALE
N
0 1 2 3 4 5
km

Figure 6. *Geological map of the Lushs Bight Group after Kean (1984) and Jenner et al. (1988) showing the geochemical affiliation of the volcanic rocks; large solid squares are volcanogenic sulphide deposits, and location indicator in legend refers to extended REE plots on Figure 7.*

Group, LREE-enriched tholeiites have clear non-arc signatures (Figures 10 and 11).

The data presented here are consistent with the paleotectonic models for the Betts Cove Complex presented by Coish *et al.* (1982). The ophiolite can reasonably be interpreted to have formed in a supra-subduction environment during rifting of an island arc. The rifting event is, in this interpretation, recorded by the boninite and arc-tholeiite volcanic rocks. Initiation of partial melting of dominantly normal mantle sources (i.e., not influenced by the subducting slab) is recorded by the MORB-like lavas in the Betts Cove Complex and by OIB-like lavas in the overlying Snooks Arm Group.

Pillow lavas in the Betts Cove Complex are host to a number of VMS deposits, the best known of which are the Tilt Cove and the Betts Cove deposits at the north and south ends of the complex, respectively (Figure 9). The former produced more than 8 mt of Cu–Zn ore and is the largest ophiolite-hosted VMS deposit in the Appalachians. In addition, there are a number of smaller deposits along strike between the two former producers (Strong and Saunders, 1988). The VMS deposits in this ophiolite, as elsewhere, show a close spatial association with the boninitic volcanic rocks. Strong and Saunders (1988) noted this association at Tilt Cove, suggesting that it resulted from metal and sulphur mass transfer from metalliferous sediments on the subducting slab into the mantle wedge, the localization of metals in immiscible sulphide droplets in the boninitic magmas, and their subsequent leaching and transport by hydrothermal fluids onto the sea floor.

The Pacquet Harbour Group

The Pacquet Harbour Group comprises polydeformed sedimentary and volcanic rocks that have been metamorphosed in the upper greenschist to lower amphibolite facies. Volcanic rocks comprise mainly pillow lavas and mafic schist that may have a tuffaceous protolith. Rhyolite is a

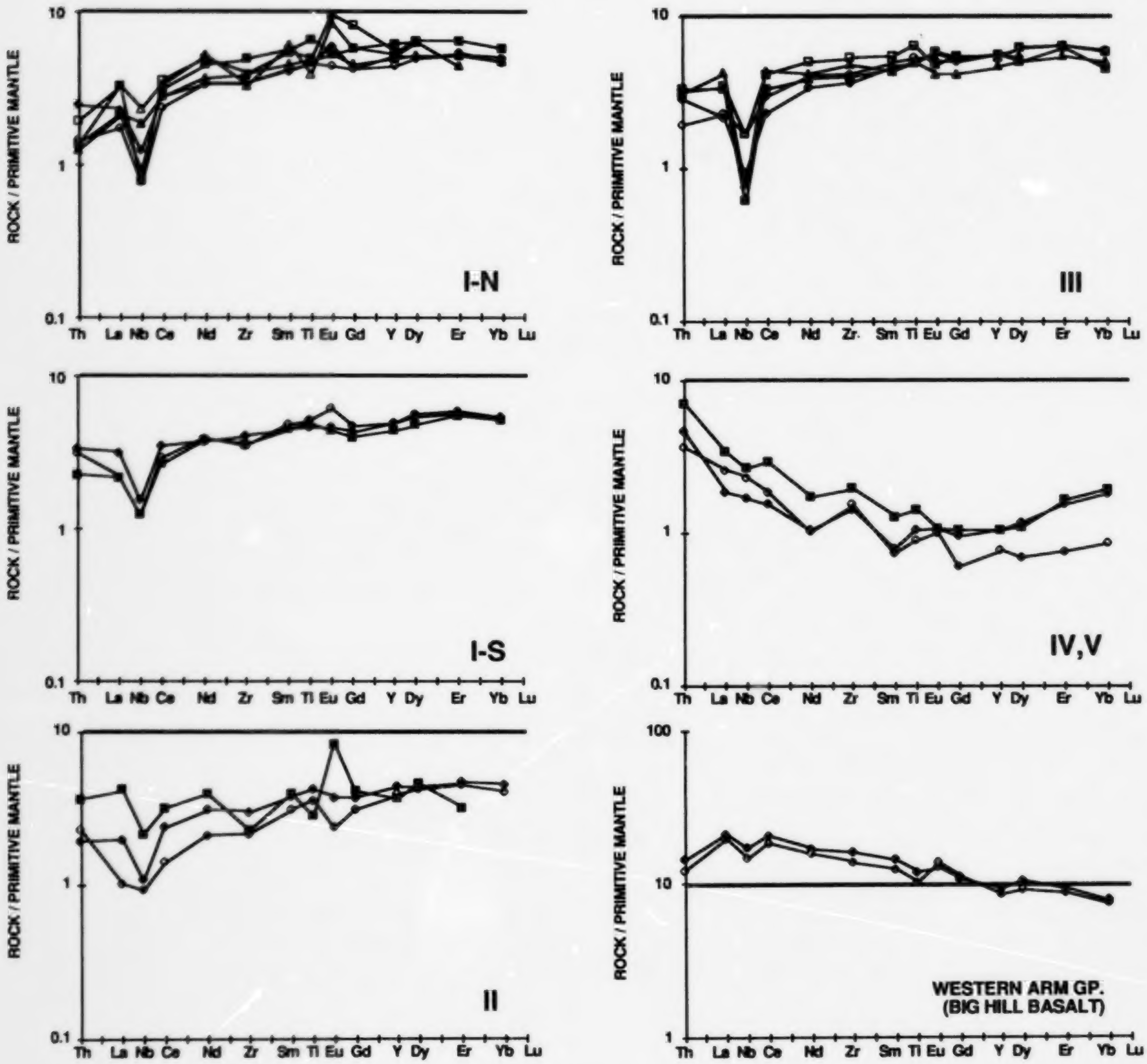

Figure 7. *Extended REE plots for volcanic rocks in the Lushs Bight Group: Subdivision described in text.*

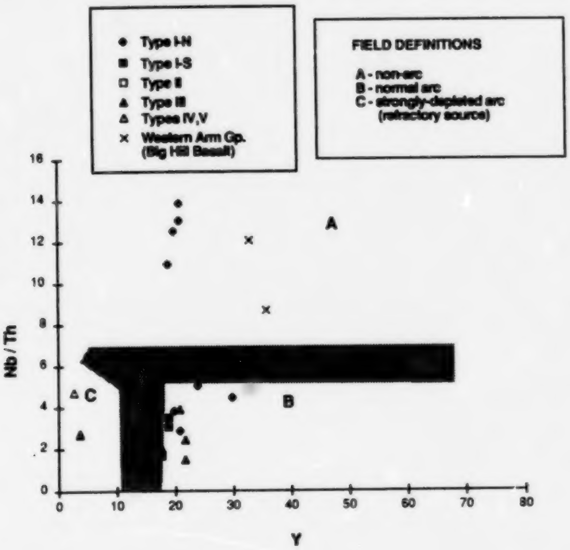

Figure 8. *Nb/Th versus Y plot for Lushs Bight Group rocks. Symbols are keyed to subdivisions illustrated in Figure 7.*

relatively minor constituent, occurring mainly in a circular outcrop area near the centre of the group (Figure 12). There is considerable uncertainty as to the internal stratigraphy of the group because of poor exposure and structural complexity (Hibbard, 1983).

Gale (1971, 1973) was the first to recognize that the Pacquet Harbour Group is geochemically bipartite, comprising tholeiitic basalts and a suite of highly magnesian, incompatible element-depleted volcanic rocks, which he termed 'basaltic komatiites'. Hibbard (1983) reported additional analyses of both rock types and remarked upon the similarity of the magnesian rocks to boninites. The trace-element data presented here, clearly show the close affinity of these rocks to boninites, with highly depleted incompatible trace-element concentrations and concave-upward REE patterns (Figure 13). One sample (termed 'depleted arc tholeiite' on Figure 13) is significantly incompatible element-depleted but apparently not a boninite as it does not have the characteristic concave-upward REE pattern. It plots in the field of overlap on Figure 14. Most tholeiites in the Pacquet Harbour Group do not carry the arc signature (Figure 13). The geochemical association is very similar to that in the less structurally disrupted and metamorphosed Betts Cove Complex and Hibbard's (1983) conclusion is accepted by us, in that the two sequences are probably related.

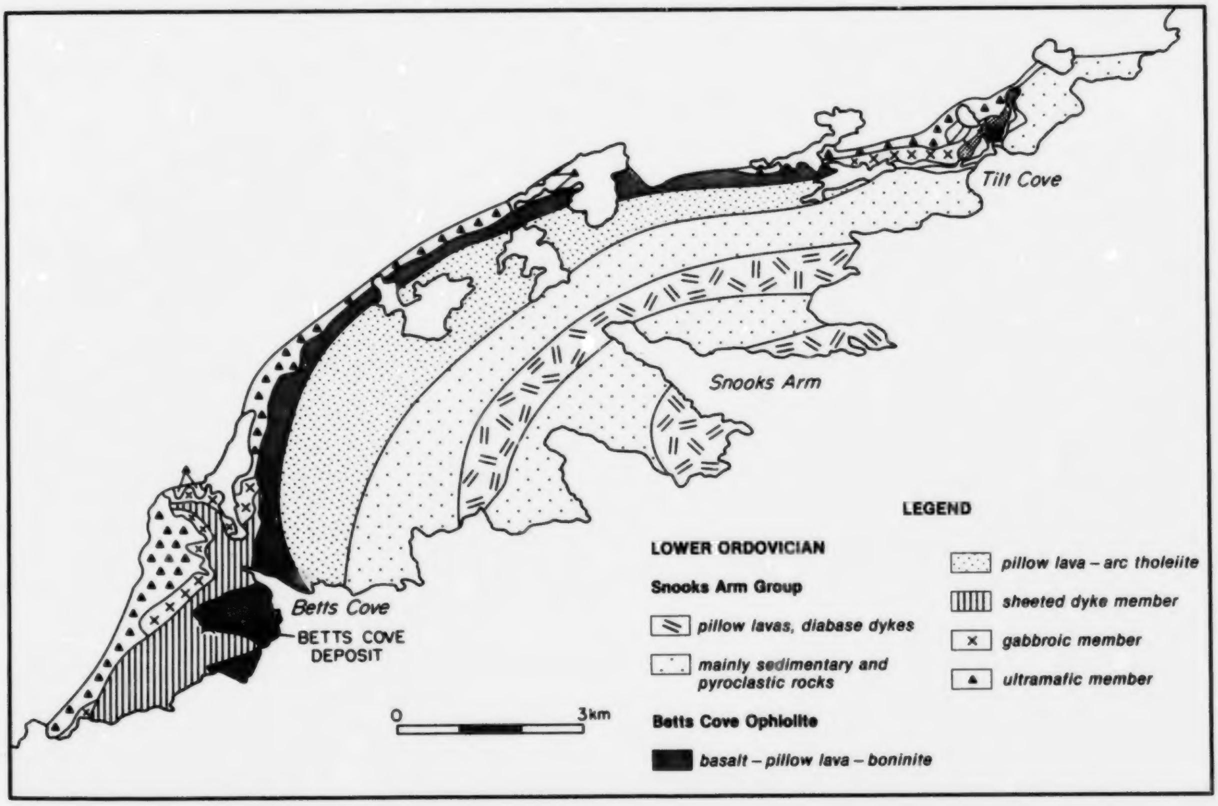
Tilt Cove
Snooks Arm
Betts Cove
BETTS COVE DEPOSIT
0
3km
LEGEND
LOWER ORDOVICIAN
Snooks Arm Group
pillow lavas, diabase dykes
mainly sedimentary and pyroclastic rocks
Betts Cove Ophiolite
basalt – pillow lava – boninite
pillow lava – arc tholeiite
sheeted dyke member
gabbroic member
ultramafic member

Figure 9. *Geological map of the Betts Cove area showing the geochemical affiliation of volcanic rocks in the Betts Cove Complex and the Snooks Arm Group; solid squares are volcanogenic sulphide deposits.*

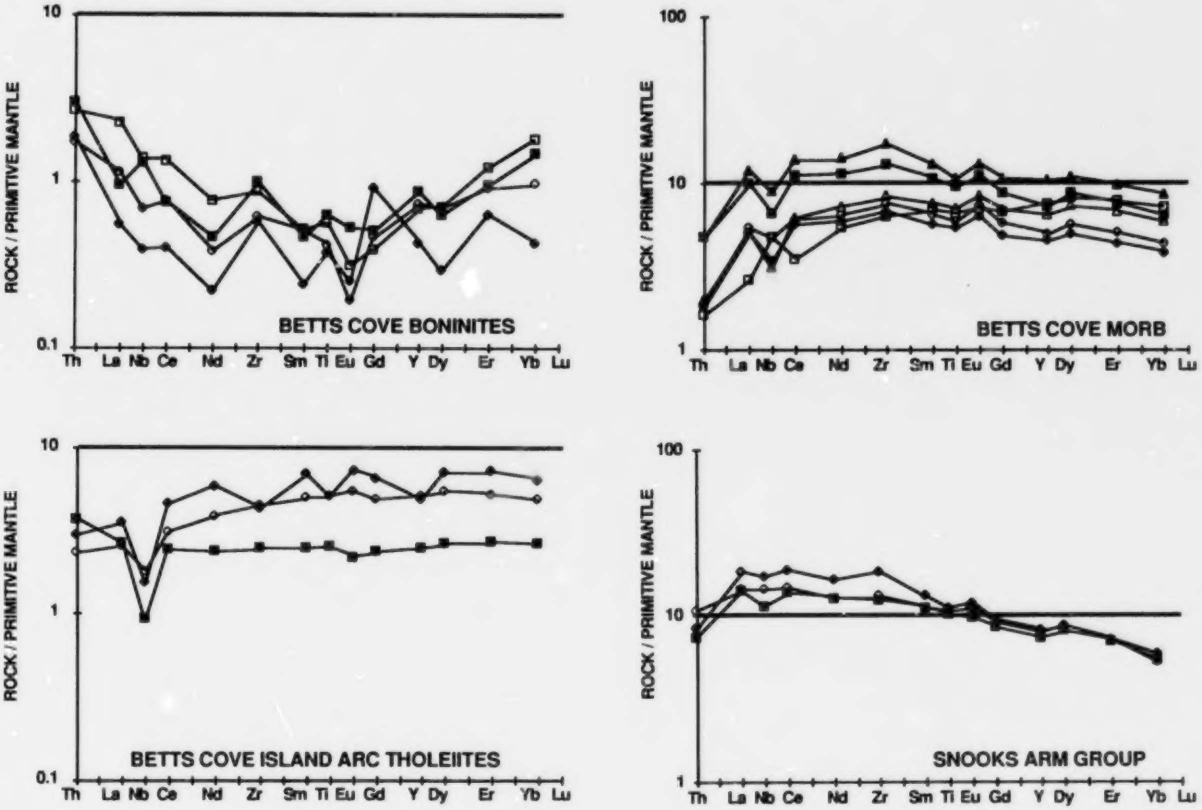

Figure 10. *Extended REE plots for volcanic rocks in the Betts Cove Complex and the Snooks Arm Group.*

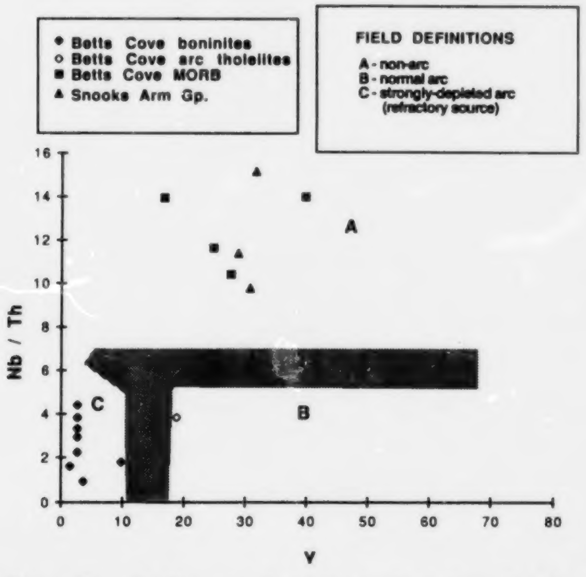

Figure 11. *Nb/Th versus Y plot for the Betts Cove Complex and the Snooks Arm Group. Symbols are keyed to subdivisions illustrated in Figure 10.*

There are five VMS deposits in the Pacquet Harbour Group, that together make up the Consolidated Rambler Mines camp (Tuach and Kennedy, 1978; Hibbard, 1983; Tuach, 1988). Three of these (Rambler, Ming, and the newly-discovered Ming West) are stratiform massive sulphides whereas the other two are footwall stockwork systems. The VMS deposits, although closely associated with the Rambler rhyolite (Tuach, 1988), are stratigraphically associated with the boninitic lavas, near the contact with the overlying tholeiites. The magnesian basalt−tholeiite transition, as in the case of the Betts Cove Complex, probably represents an arc-rifting to back-arc basin transition and as in the Betts Cove example, the VMS deposits appear to be associated with this tectonic transition.

The Tally Pond and Tulks Hill Volcanic Belts

The Tally Pond and Tulks Hill volcanic belts comprise Cambrian to Lower Ordovician mafic and felsic volcanic rocks that occupy the western and eastern margins, of the Victoria Lake Group, respectively (Figure 15). They include considerably greater proportions of felsic volcanic rocks than the sequences in Notre Dame Bay. Previous workers have included the volcanic sequence near the Victoria Mine as part

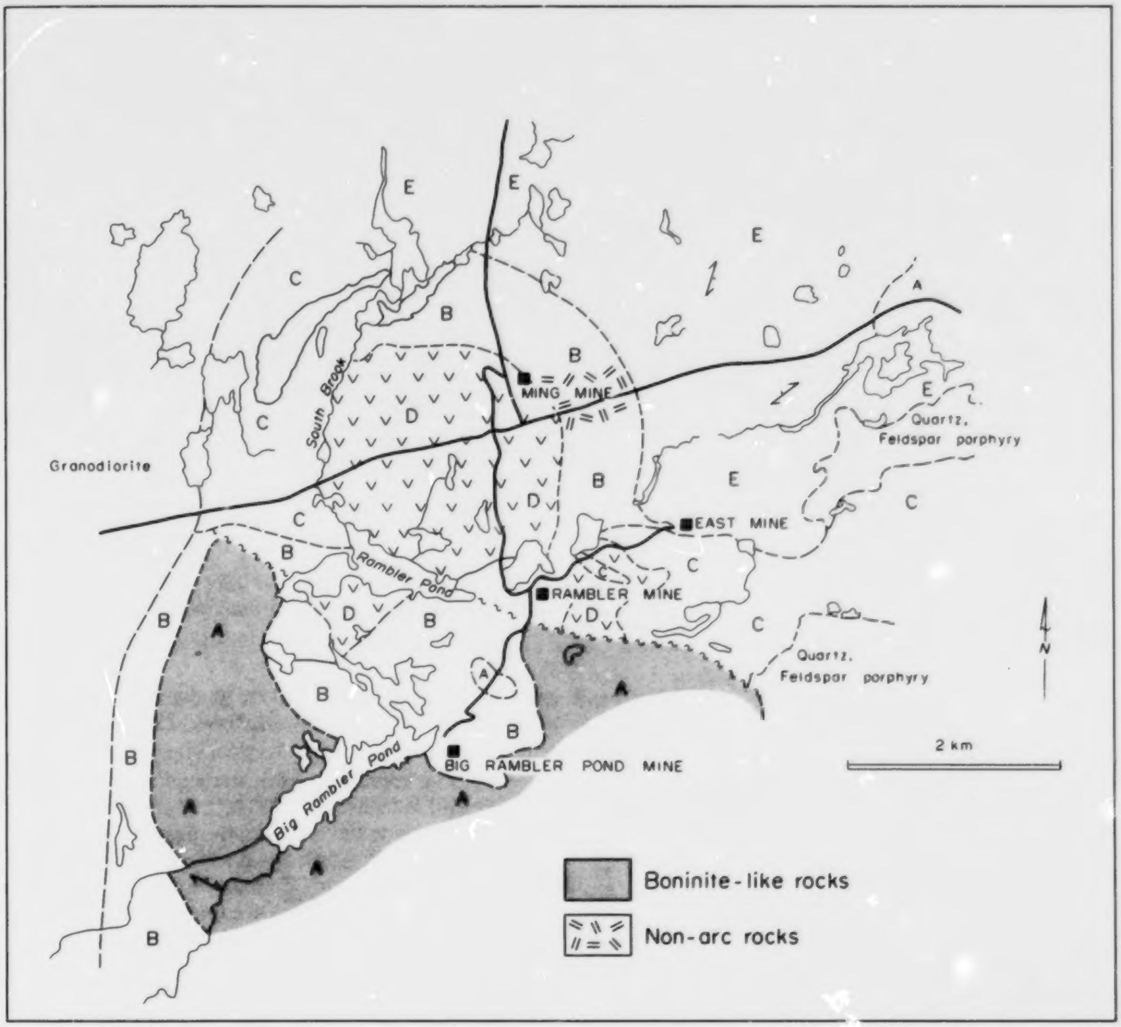
E
E
E
C
B
B
South Brook
D
MING MINE
Granodiorite
C
C
B
D
EAST MINE
E
Quartz, Feldspar porphyry
C
Rambler Pond
RAMBLER MINE
D
B
C
B
A
Quartz, Feldspar porphyry
N
2 km
B
B
A
B
BIG RAMBLER POND MINE
Big Rambler Pond
A
A
A
B
A
Boninite-like rocks
Non-arc rocks

Figure 12. *Geological map of the Rambler Mine area after Tuach (1988) showing the geochemical affiliation of volcanic rocks in the southwestern part of the Pacquet Harbour Group: A, mafic flow rocks; B, mafic volcaniclastic rocks; C, mixed mafic and silicic rocks; D, silicic pyroclastic rocks; E, clastic sediments.*

of the Tulks Hill volcanic belt (e.g. Kean *et al.*, 1981). However, following Kean and Evans (1988b) and Swinden *et al.* (1988), the Victoria Mine sequence is considered to be younger than the main part of the belt and is not included in the following discussions.

The two volcanic belts together make up approximately 60 percent of the Victoria Lake Group, the remainder being a large epiclastic basin in the central and northern parts of the group (Figure 15). Both volcanic belts are characterised by laterally extensive, predominantly felsic, pyroclastic rocks. Mafic pillow lavas and massive flows are common in the Tally Pond volcanics. In the Tulks Hill volcanics, mafic rocks are dominantly tuffaceous, consisting of aquagene tuff, lapilli tuff, agglomerate and breccia. Exposure is generally poor throughout the Victoria Lake Group and contact relationships can seldom be demonstrated on the ground.

Both sequences are prolific hosts to volcanogenic sulphide deposits. The Tulks Hill volcanics contain major massive sulphide deposits at Tulks Hill and Tulks East (Barbour and Thurlow, 1982; Kean and Evans, 1986; Evans and Kean, 1987), and several other minor prospects. The Tally Pond volcanics host the recently discovered Boundary and Duck Pond deposits (Kean, 1985; MacKenzie, 1988; MacKenzie *et al.*, 1988) as well as a number of minor prospects.

Although our data for this particular area are very preliminary, they do suggest a relationship between host-rock geochemistry and volcanogenic sulphide deposits. In the Tulks Hill belt, three geochemical subdivisions of mafic volcanic rocks are recognized, from east to west: 1) several outcrop areas of arc-tholeiites (probably including all of the Harmsworth Steady volcanics, Figure 15) that range from

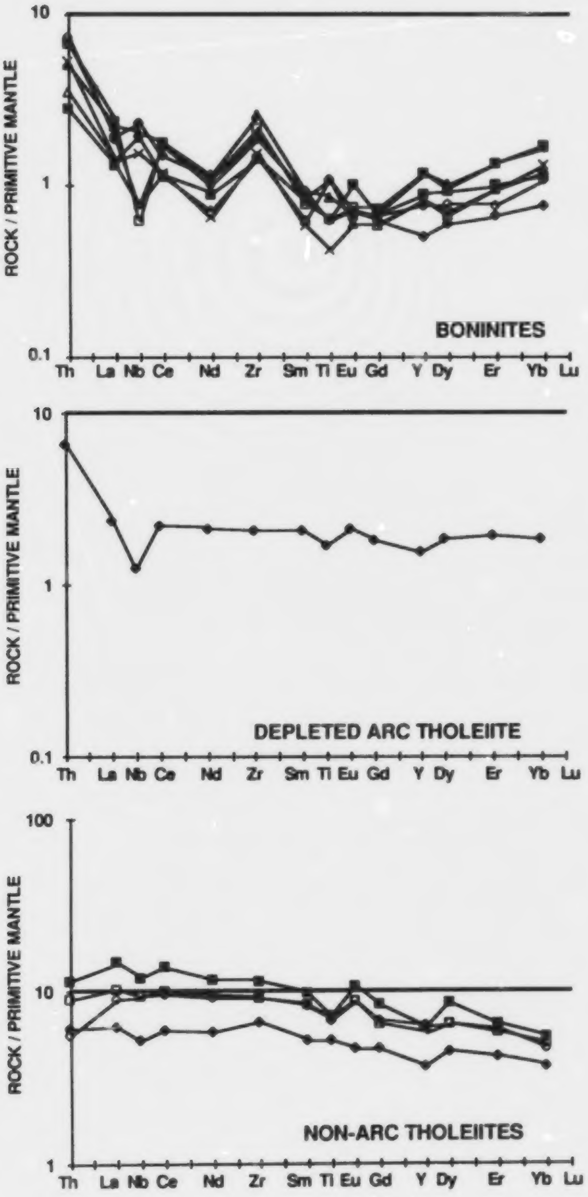

Figure 13. *Extended REE plots for volcanic rocks in the Pacquet Harbour Group.*

slightly to strongly LREE-enriched (Figure 16); 2) a sequence of dominantly felsic volcanics that locally are interbedded with strongly depleted arc-tholeiites; and 3) a sequence of non-arc tholeiites (the upper basalts). All of the volcanogenic sulphides in this belt are hosted by felsic volcanic rocks. However, analyses of relatively unaltered pillow lavas, massive flows and mafic dykes from outcrop and drill core near many of these deposits, have shown that there is commonly a spatial association beween VMS deposits and highly depleted arc-tholeiite volcanic and/or subvolcanic rocks. Both the eastern arc-tholeiite sequences and the western upper basalts of non-arc origin are apparently barren.

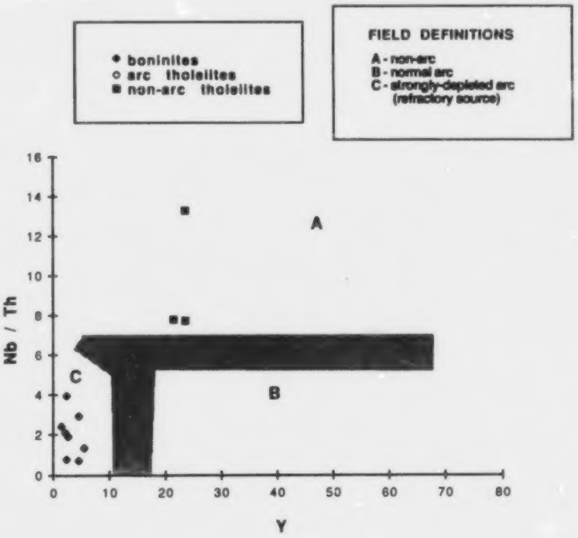

Figure 14. *Nb/Th versus Y plot for the Pacquet Harbour Group. Symbols are keyed to subdivisions illustrated in Figure 13.*

The interpretation of the geochemical relationships in the Tulks Hill volcanics is consistent with the generally west-facing stratigraphic succession proposed by Kean and Jayasinghe (1980), which appears to record the transition from island arc to back arc. Because of the lack of exposure in the area, it is not possible to document this interpretation with the same confidence as in the better exposed sequences in Notre Dame Bay. It is possible that the contacts are structural and the different basalt types are not genetically related. However, the presence of highly depleted arc-tholeiites near the major sulphide deposts (Figure 17) does suggest that at least these sequences probably formed in an arc-rifting environment.

In the Tally Pond area, the data to date are even less definitive. Rocks in the Lake Ambrose area are mainly arc-tholeiites having REE patterns varying from relatively flat to somewhat LREE-enriched (Figure 18). As yet, there are no detailed data from volcanic rocks near the Duck Pond and Tally Pond VMS deposits. In the northeastern part of the volcanic belt, highly depleted rocks (Figure 19) with concave-upward REE patterns have been identified in outcrops near Sandy Lake (Figure 15). These rocks may be boninites and are spatially associated with the recently discovered Old Sandy Road massive sulphide prospect (Kean and Evans, 1988c). This prospect was found during reconaissance prospecting following initial recognition that this rock type was present in the area. Based on whole rock geochemistry, the Sandy Lake belt would appear to be a highly prospective area for volcanogenic sulphide exploration.

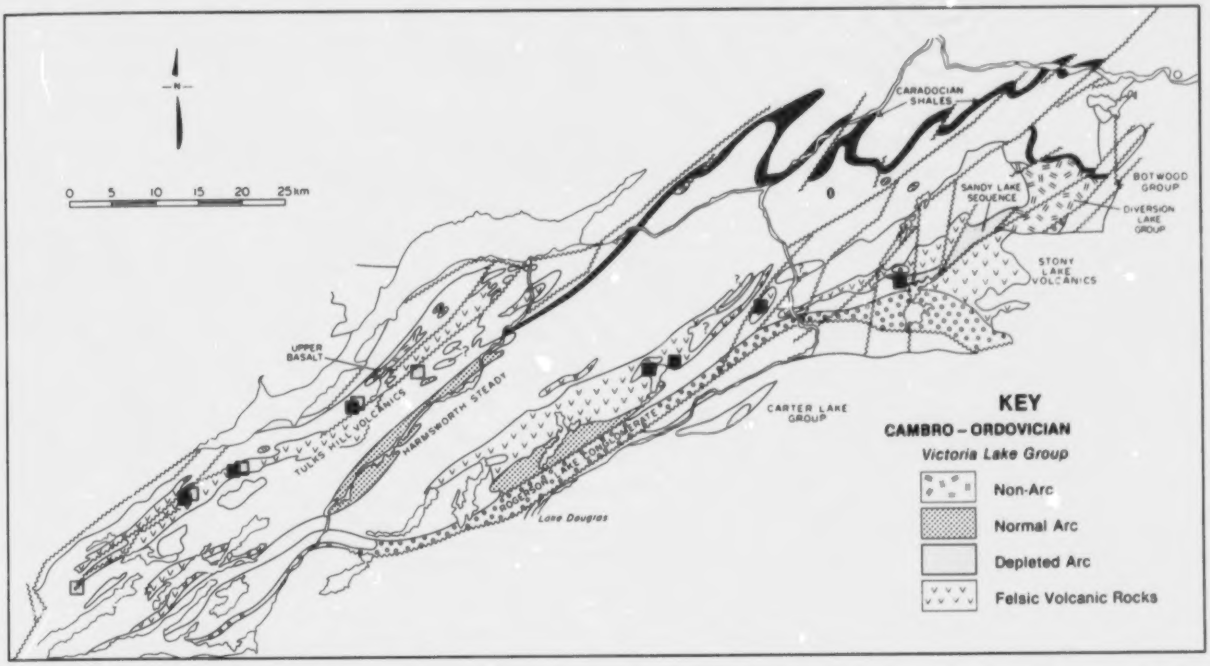
N
0 5 10 15 20 25 km
CARADOCIAN SHALES
SANDY LAKE SEQUENCE
BOTWOOD GROUP
DIVERSION LAKE GROUP
STONY LAKE VOLCANICS
UPPER BASALT
TULKS HILL VOLCANICS
HARMSWORTH STEADY
CARTER LAKE GROUP
ROGERSON LAKE CONGLOMERATE
Lake Douglas
KEY
CAMBRO – ORDOVICIAN
Victoria Lake Group
Non-Arc
Normal Arc
Depleted Arc
Felsic Volcanic Rocks

Figure 15. *Geological map of the Victoria Lake Group after Kean and Evans (1988c) showing the geochemical affiliation of the volcanic rocks in the Tulks Hill and Tally Pond volcanic belts; solid squares are volcanogenic sulphide deposits and grey squares are pillow lava outcrops too small to show at this scale.*

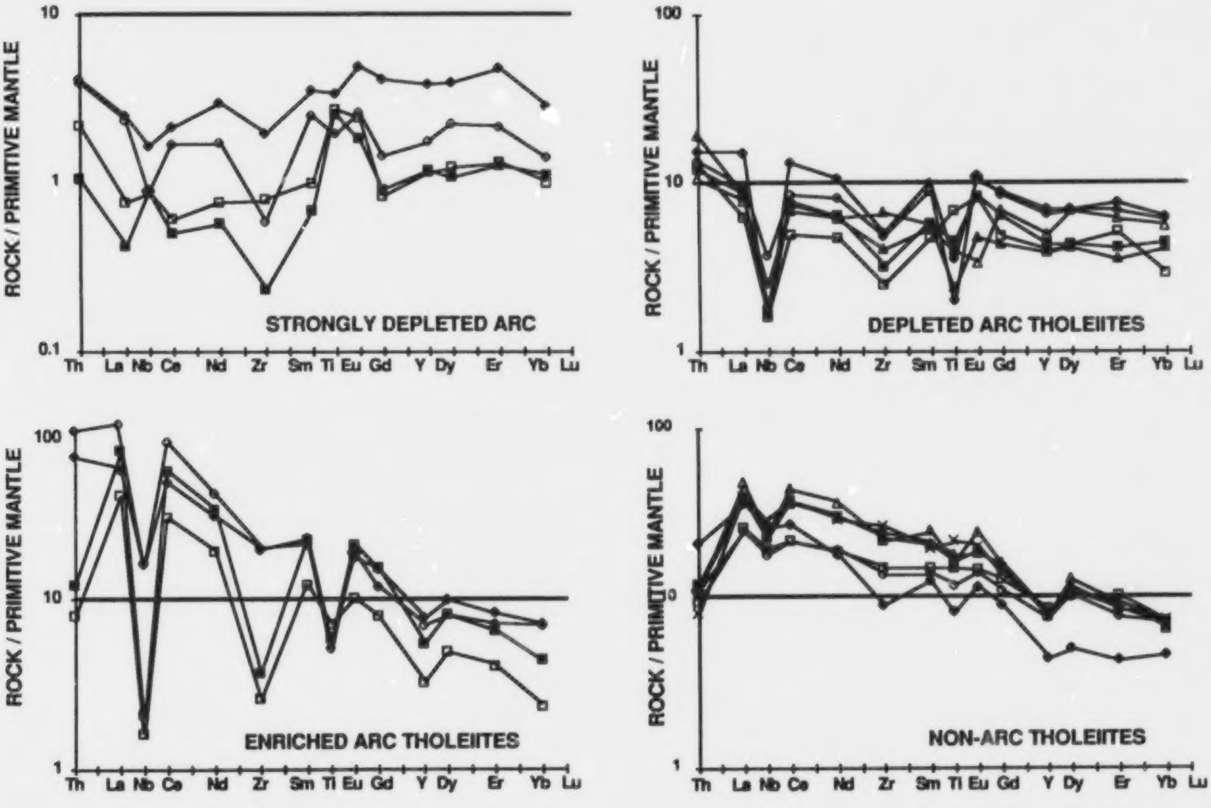

Figure 16. *Extended REE plots for volcanic rocks in the Tulks Hill volcanic belt.*

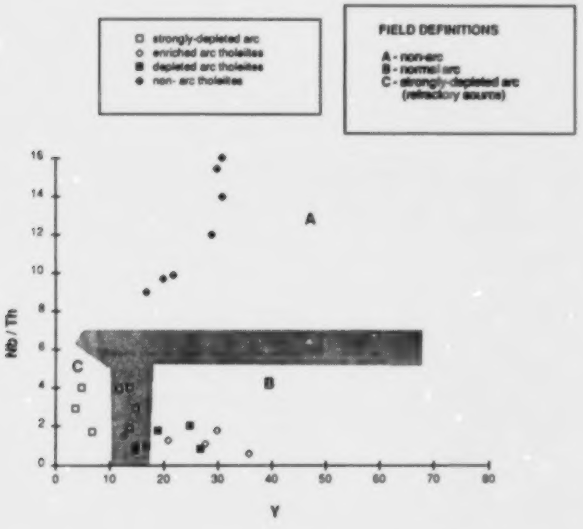

Figure 17. *Nb/Th versus Y plot for the Tulks Hill volcanic belt. Symbols are keyed to subdivisions illustrated in Figure 16.*

The data for the Diversion Lake group at the extreme northeastern end of the belt suggest that they are dominantly LREE-enriched non-arc tholeiites (Figure 18) and unlikely to be prospective.

WHERE ARE THE MASSIVE SULPHIDES?

In the five sequences studied, there are some obvious consistencies in the settings of volcanogenic sulphide deposition, as well as some significant differences. Some points of comparison are:

1) *Geochemical associations in the volcanic sequences:* A geochemical association that includes both rocks with arc signatures and those with non-arc signatures can be recognized in all areas. In two areas (the Wild Bight Group and the Betts Cove Complex), the field relationships are sufficiently well established to show that there is a geochemical progression upward through the stratigraphy from island-arc rocks to non-arc rocks and these sequences are interpreted to record the rifting of an island arc and the subsequent establishment of a back-arc basin. This interpretation serves as a model for other less well exposed or more structurally-disrupted sequences, where the elements of the model appear to be present but cannot be unequivocally related to each other, stratigraphically. The Lushs Bight–Western Arm groups, the Pacquet Harbour Group and the Tally Pond and Tulks Hill volcanics all contain elements of a similar tectonic transition. Because the units are not all the same age (Swinden *et al.*, 1988), island-arc rifting probably occurred more than once during the long history of Iapetus. By analogy with modern environments such as the southwest Pacific Ocean, it may have been a relatively common event.

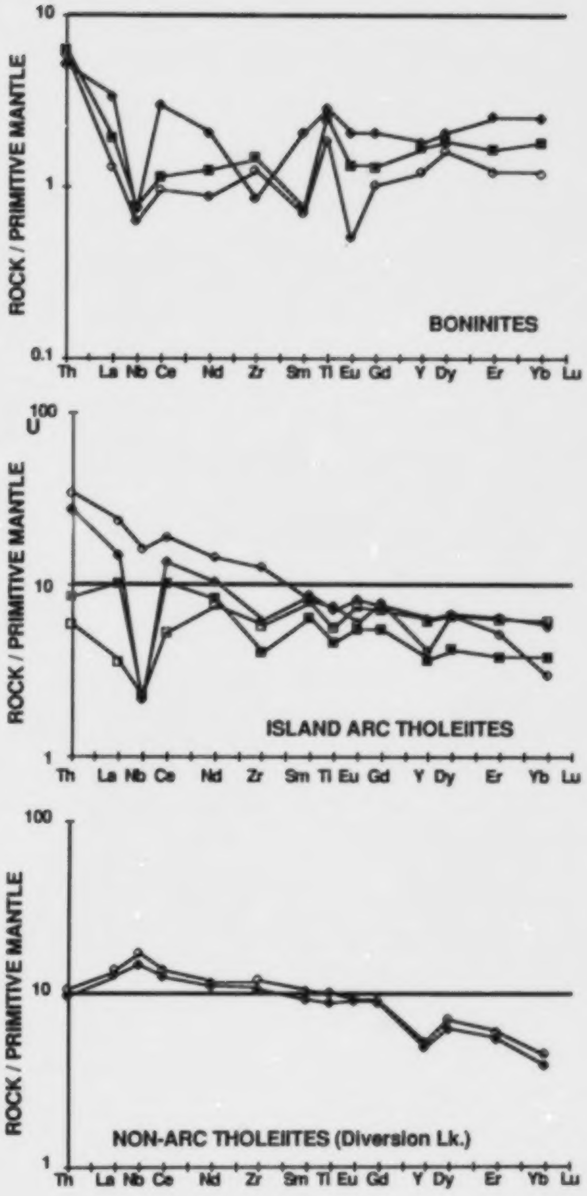

Figure 18. *Extended REE plots for volcanic rocks in the Tally Pond volcanics.*

2) *Geochemical association of VMS deposits:* Volcanogenic sulphide deposits are generally associated with rocks having a clear island-arc geochemical signature. From the preliminary studies reported here, and from studies of other sequences reported elsewhere (see review, Swinden, 1987), we are aware of only one exception, this being the Great Burnt Lake deposit in south-central Newfoundland (Swinden, 1988).

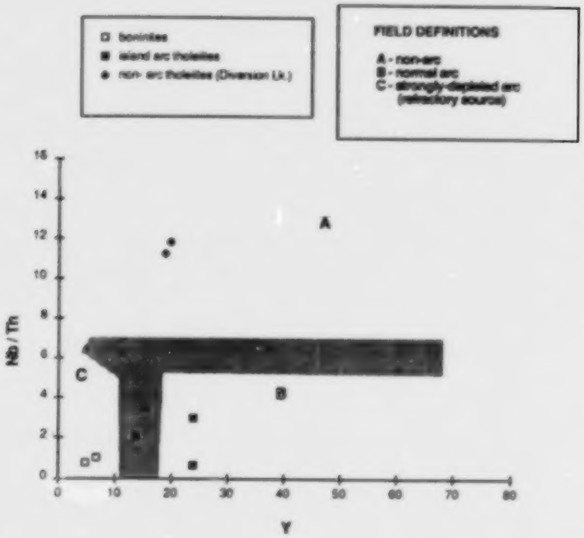

Figure 19. *Nb/Th versus Y plot. Symbols are keyed to subdivisions illustrated in Figure 18.*

3) *Paleotectonic setting of mineralization and association with refractory source melts:* It appears that in the central Newfoundland VMS districts, mineralization took place mainly during the arc-rifting stage. In all areas, volcanogenic sulphide deposits are spatially associated with highly incompatible element-depleted mafic lavas that must represent the product of complex hydrous remelting of refractory mantle sources. This association is not always intimate; in many cases, deposits are hosted by felsic volcanics (e.g., Point Leamington, and most deposits in the Tulks Hill and Tally Pond volcanics) or by normal arc-tholeiites (e.g., many deposits in the Lushs Bight Group). However, there are usually highly incompatible element-depleted lavas (i.e., refractory source melts) nearby. It would appear that although the presence of this type of magma is important to the mineralizing process, it is not always the host of the associated mineral deposits.

Clearly, the close association of these peculiar depleted rocks and volcanogenic sulphides implies a genetic connection. However, the exact nature of this genetic connection is by no means clear. A brief consideration of the contrasts between different mineralized settings may help constrain or eliminate some interpretations by eliminating from consideration processes that can not be demonstrated in all areas. These contrasts include:

1) *Stratigraphy of the mineralized sequences:* The Betts Cove Complex is an ophiolite. The Pacquet Harbour and Lushs Bight groups may also be derived from an ophiolite, as suggested by previous workers, although this cannot be unequivocally demonstrated. The Wild Bight and Victoria Lake groups are thick sequences of volcanic and volcaniclastic rocks. In the former, mafic rocks dominate the volcanic successions, with felsic volcanics restricted to small accumulations that make up less than 5 percent of the volcanics. In the latter, laterally extensive felsic volcanic sheets make up more than 35 percent of the volcanic succession.

2) *VMS host rocks:* Although generally associated with highly depleted mafic volcanic rocks, massive sulphides may be hosted by a variety of lithologies including felsic volcanics, normal arc-tholeiites and depleted rocks (see above).

3) *Spatial association with rhyolites:* Volcanogenic sulphides are closely associated with felsic volcanics in many areas (Wild Bight, Pacquet Harbour and Victoria Lake groups and the western part of the Lushs Bight Group). However, in the Betts Cove Complex and in the central and eastern parts of the Lushs Bight Group, there are no associated felsic volcanics.

4) *Diverse geochemical characteristics of refractory magmas:* The highly-depleted mafic volcanic rocks are not all of uniform geochemical characteristics. In the Betts Cove Complex and in the Pacquet Harbour Group, all have concave-upward REE patterns, positive Zr and negative Ti anomalies (on a normalized basis, see Figures 10 and 13) similar to boninites (similar rocks occur in the Sandy Lake belt of the Tally Pond volcanics). In the Wild Bight Group, they are tholeiitic, having peculiar and distinctive flat REE patterns with a step-up between Sm and Eu, negative Zr and positive Ti anomalies (Figure 4). In these two rock types, Zr and Ti have been decoupled from the REE during petrogenesis but apparently in a different fashion. The Lushs Bight Group contains both highly depleted tholeiitic rocks and boninite-like rocks, and both are associated with volcanogenic sulphide deposits.

GENETIC ASPECTS

In all the cases examined, VMS deposition occurred during a period of magmatic activity that is best interpreted as rifting of an island arc. This appears to be true irrespective of the nature of the host sequence, i.e., whether it is an ophiolite or a volcanic–volcaniclastic sequence. As previous authors have noted, rifting events of any kind provide an uncommonly good opportunity for volcanogenic mineralization (Sillitoe, 1982; Cathles *et al.*, 1983). This is particularly true in oceanic environments, where the rifting provides deep open fractures that enhance the permeability of the substrate, and promote deep access of sea water. The rifting is generally accompanied by crustal thinning, resulting in enhanced heat flow that may be an order of magnitude greater than in normal arc settings (Cathles *et al.*, 1983). The combination of high heat flow and enhanced permeability provides an ideal opportunity for enhanced hydrothermal circulation with concomitant metal transport and sulphide deposition (Cathles, 1983). The presence of volcanogenic mineralization in this setting in central Newfoundland is, therefore, entirely consistent with current concepts of island-arc metallogenesis.

In the sequences that have been studied, mineralization is spatially associated with rhyolitic volcanism. However, this does not seem to be a necessary condition in this environment, as deposits in the Lushs Bight Group and the Betts Cove Complex do not have associated rhyolite. Conversely, there are major accumulations of rhyolite in the Wild Bight Group that have no associated VMS deposits. This is in accord with the conclusions of Cathles (1978), Urabe and Sato (1978), Ohmoto (1978) and Cathles *et al.* (1983) who have suggested that rhyolite domes like those commonly associated with ore deposits in the Kuroko District, Japan, are probably too small to have had a cause and effect relationship with the VMS deposits. However, the formation of rhyolite through melting in the basal arc crust would undoubtedly be promoted by the anomalous heat flow and fluid circulation during the arc-rifting stage and, therefore, the association of rhyolites with VMS deposits is probably not fortuitous.

The unifying characteristic in all mineralized rifted arc settings in central Newfoundland is the presence of basalts that are best interpreted as hydrous melts of refractory sources. The tectonic and magmatic conditions that allowed these rocks to be erupted at surface are apparently both necessary and sufficient for VMS deposition in these settings.

This implies a genetic link between the refractory source melts and the VMS deposits. This genetic link may be the favourable tectonic environment for fluid circulation coupled with increased heat flow during ascent of these melts. Because the refractory source melts require so much heat of fusion, they must rise to the surface rapidly (or they would freeze on the way up), and probably do not pool in shallow magma chambers for extended periods. They probably make use of deep, throughgoing fractures related to the rifting to make their quick ascent to surface. The opening of fractures in this environment would promote additional heat flow irrespective of the rise of the hotter magmas. Because the bulk fluid/rock ratio in a sub-seafloor hydrothermal convection cell is dependant upon heat flux (among other things), this would result in increased fluid circulation, an expanded volume of rock that would see fluid at a temperature appropriate for leaching metals, and a corresponding increase in the amount of metal available to the system for eventual deposition at the surface. The fracturing related to rifting would also have the effect of focussing fluid flow on the up-flow side of the hydrothermal cell, allowing a greater volume of metal-charged fluid to reach the subsurface with the metals intact.

The overall effect would be an enhanced environment for the generation of VMS deposits. Geological intervals in which these refractory source melts occur are places where the chances of discovering a VMS deposit appear to be considerably enhanced. The combination of tectonic and hydrothermal circumstances that are extant during arc rifting, and particularly during eruption of very hot refractory source melts, are the necessary and sufficient conditions that make these environments so prolific.

This interpretation, is at odds with the models of Strong and Saunders (1988) for the deposits at Tilt Cove. Although the highly depleted melts may locally contain sulphide droplets, the connection between these sulphides and subducted metalliferous sediments seems tenuous and in most areas of the present study, there does not appear to be a sufficient volume of depleted mafic volcanic rock to have provided the metals to the associated massive sulphide deposits (this is particularly true in the Wild Bight Group). Although sulphides in the depleted magmas may contribute some metals to the hydrothermal system, there would seem to be no need to rely on them solely for enhanced metal concentrations in the hydrothermal cells. The increased heat in the rifted-arc environment, driving larger and hotter hydrothermal cells, is probably sufficient to account for the increased VMS production in this setting.

SUGGESTIONS FOR USE IN EXPLORATION

The presence of refractory source melts in oceanic volcanic sequences may be useful as a regional exploration tool for volcanogenic sulphides. We offer the following conclusions based upon our preliminary results:

1) The prime utility of whole rock geochemical studies is likely to be in the early stages of exploration in a volcanic belt. Geochemical studies of mafic volcanic rocks provide a rapid means of discriminating areas of high potential from those of lower potential within a given volcanic belt, or as a means of assigning exploration priorities when looking at more than one volcanic belt.

2) In order to make the best use of this approach, careful sampling and sample preparation techniques are required. It is essential that the rocks being sampled are correctly identified. Taking indiscriminant samples from a heterogeneous assemblage of mafic igneous rocks with no evidence as to whether they represent flows or intrusives will inevitably lead to considerable geological noise in the data and may yield incorrect interpretations. As a general rule, if the field evidence is not clear as to the origin of the rock, it is better not to sample it. Pillow lavas are the prefered sample lithology. In particular, the crystalline interiors of the pillows are the preferred sample medium as they are less likely to have experienced mass transport during alteration.

3) Analytical techniques must be chosen with care and closely monitored for accuracy and precision. In particular, the depleted volcanic rocks have trace-element concentrations that are well below reliable determination by routine analysis in commercial laboratories. To make most effective use of this technique, confidence in the data at very low detection limits is necessary.

4) Volcanic sequences in which the rocks have a non-arc signature can probably be assigned a very low exploration priority at this stage.

5) Within sequences where the rocks have an arc signature, the presence of highly depleted lavas (i.e., that plot in Field 'A' on the Nb/Th versus Y diagram) is a positive exploration indicator. Note that the VMS deposits may not necessarily be hosted by the depleted lavas. For example, the hydrothermal fluids may have vented in an area somewhat removed from the eruptive centres and lie upon a different volcanic rock. Alternatively, there may be a time gap between eruption of the depleted lavas and venting of the hydrothermal systems, with other volcanic or sedimentary rocks having been deposited in the interim. The important point is that the presence of these rocks appears to be a better than average indicator that the mineralizing process may have occurred, and that economic deposits may profitably be sought in the vicinity.

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RECONNAISSANCE ASSESSMENT OF POTENTIAL BEDROCK AGGREGATE

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ABSTRACT

In 1988, a reconnaissance assessment of the aggregate potential of bedrock in the Clarenville and Gander areas was undertaken. Natural aggregates (sand, gravel and crushed stone) are nonrenewable mineral commodities that are being continuously depleted by the rapidly increaseing demand of the construction industry for aggregate use in cement, concrete, asphalt, railway ballast, armour stone, gravel roads and fill material.

A total of 294 sites were visited and 243 samples were collected. Of the 294 sites visited, 187 sites were identified as showing potential for bedrock aggregate; 62 sites were identified as marginal aggregate sites and the remaining 45 sites were considered to be of poor quality for aggregate use.

Geotechnical properties of the bedrock, including geological structures, deleterious substances, abrasion, soundness, petrographic analysis for potential alkali-reactivity and petrographic number are presently under study.

INTRODUCTION

During the 1988 field season, an evaluation of bedrock aggregate potential in the Clarenville and Gander areas was undertaken. This assessment was conducted to determine the quantity and quality of bedrock aggregate for industrial use in the area. This inventory should help to eliminate the use of poor quality rock (e.g., alkali-reactive rocks that react destructively in concrete, and friable rock as a form of armour stone) (Plates 1 and 2).

Plate 1. *Bridge deterioration that is believed to be caused by alkali-reactive rocks.*

The field area covers sixteen 1:50,000 map sheets (Figure 1) and because of time and staff limitations, only road traverses were carried out.

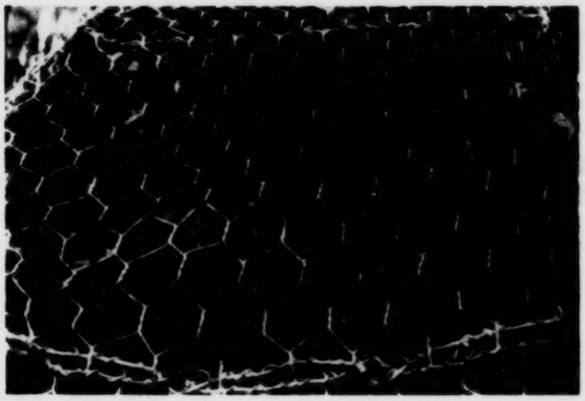

Plate 2. *Friable shale used as a form of armour stone.*

FIELD WORK

Field work consisted of a detailed examination of all rock quarries, road-cuts and natural outcrops along all highways and side roads in the study area. Each site investigation entailed rock identification, representative sampling, determination of overburden type and thickness, and the recording of all geological features present; these site investigations are described below.

Rock Identification

Rocks were identified according to their genetic origin as being either sedimentary (a rock formed by the deposition

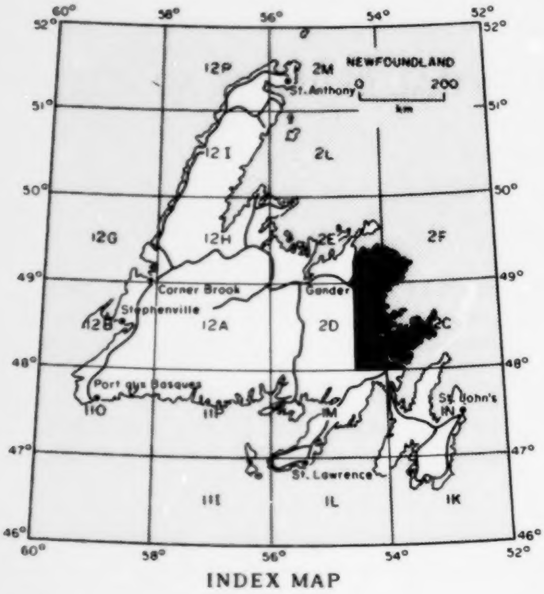

Figure 1. *Location of study area.*

of sediment), igneous (a rock or mineral that solidified from molten or partly molten material), or metamorphic (rocks that have been changed or deformed by high temperature and pressure within the earth's crust). Identification was usually made on site with the aid of a hand lens, but a later more detailed laboratory examination using a petrographic microscope was made, to enable a more precise and accurate rock classification.

Representative Sampling

Sampling is the most critical aspect of the investigation because unless a sample is representative of the site, any testing or analysis undertaken would be clearly, of no value.

A minimum of one representative sample (1 kg) was collected from each quarry or sample site. However, where the quarry or sample site was large (greater than 300 m^2), had a number of different rock types exhibiting varying degrees of weathering and where veining or alteration zones were observed, between 2 and 5 samples were commonly collected. In the latter instances, much larger samples (about 50 kg) were collected at random for more extensive laboratory analyses.

Overburden Type and Thickness

The predominant type of overburden observed in the assessment areas was a veneer (< 1 m) of till (observed at about 80 percent of sites), however, locally, thicknesses ranged from 0.5 to 3.5 m. The remaining sites are thinly vegetated with shrubs and trees.

Geological Structures

The main geological structures that were studied during the field investigation included faults and dykes (Plate 3), fractures (Plate 4), joints (Plate 5) and cleavage (Plate 6). These control the breakage pattern of the bedrock and hence the cost of quarrying, by determining the amount of explosives required and the amount of waste products (such as excessively oversized or undersized blast material).

Plate 3. *Road-cut showing a mafic dyke (darker colour); dyke is also offset by a fault.*

Plate 4. *Numerous fractures in a road-cut.*

Other characteristics that are examined in the field include:

1) grain size—generally, the finer the grain size the harder the rock;

2) bedding structures—important because rocks tend to break more easily along bedding planes and laminae;

3) flow structures—in igneous rocks these tend to be planes of weakness;

Plate 5. *Vertical jointing in granite.*

Plate 6. *Metasediments, showing schistosy and prominent cleavage.*

4) mineral alignment—in metamorphic rocks these tend to be planes of weakness and thus important with regard to excessive rock breakage when blasting, crushing and during blending of crushed stone with cement or asphalt binders;

5) mineralization, veining and alteration zones were also noted because of their deleterious affect on asphalt or concrete; and

6) weathering—the degree of weathering (fresh, slight, moderate and high) is important because it affects the durability of the aggregate (i.e., the more weathered a rock, the weaker it would be and thus less durable).

GENERAL GEOLOGY

The Gander area consists primarily of Ordovician sedimentary rocks, sandstone, and shale having varying degrees of metamorphism and psammitic and pelitic rocks of the Gander Group (McGonigal, 1973), and also sedimentary rocks (shale, siltstone, sandstone, limestone) of the Davidsville Group (Kennedy and McGonigal, 1972). The area is intruded by a series of plutonic rocks (granites, granitic-gneiss) of Ordovician (Hare Bay Gneiss, Blackwood, 1982), Silurian (Cape Freels and Middle Brook granites, Strong *et al.*, 1974) and Devonian (Ragged Harbour Pluton and Deadman's Bay Granite, Strong *et al.*, 1974) age.

The Clarenville area consists of Precambrian sedimentary rocks (siltstones, sandstones, graywackes, shales and conglomerates) of the Connecting Point and Musgravetown groups (Hayes, 1948), and minor amounts of felsic to mafic volcanic rocks of the Bull Arm Formation (McCartney, 1958). Precambrian volcanic rocks (basalts, rhyolites, tuffs and breccia) of the Love Cove Group (Jenness, 1958), and minor Cambrian shales of the Adeyton Group (Jenness, 1958) were also noted.

Rock Types and Uses

Rock types in any given area that are being considered for aggregate use should initially be rated on the basis of the amount of deleterious substances present, and the petrographic number.

Deleterious substances are materials that occur in or on the rock and are capable of producing adverse effects (e.g., chemical reactions with other minerals), resulting in a deterioration of the rock or cement binder used in concrete or asphalt. Some commonly found deleterious substances include clays, organic matter, mica, iron and manganese oxide staining and cherty or fine grained siliceous material. Alteration zones, encrustations and the degree of weathering are also factors that are considered to be deleterious to the rock.

A petrographic number is calculated for each site and this measures the initial quality of material for aggregate purposes. The petrographic number is calculated by sampling 100 clasts and assigning a petrographic factor to each clast. The petrographic factor ranges from 1 (best) to 10 (worst) depending on rock type, freshness, hardness and weathering. The petrographic number is the sum of the petrographic factor for each clast and thus can range between 100 and 1000. Each rock type is given a petrographic factor of 1, 3, 6 and 10 (CSA standard A23.12) (Table 1) and a revised version (Table 2; Bragg, 1986) are factors given to different rock types that indicate the initial assessment of the rock for aggregate use. A petrographic factor of 1 indicates the highest quality, whereas a petrographic factor of 10 indicates aggregates of the lowest quality. Thus, the lower the petrographic number the higher the rock quality, which is expressed in the range of 100 to 1000 (e.g., a clean, hard unweathered granite would normally have a petrographic factor of 1, and a petrographic number of 100, whereas a friable, soft shale would have a factor of 10 and a number of 1000). The quality is usually affected by degree of weathering of the rock (Table 3). Table 3 shows the range of petrographic numbers of the different rock groups that were sampled during the field season.

Table 1. Rock type and petrographic factors

Rock Type	Classification	Factor
Carbonates (hard)	good	1
Carbonates (sandy, hard)	good	1
Sandstone (hard)	good	1
Gneiss (hard)	good	1
Quartzite (coarse grained)	good	1
Greywacke–arkose	good	1
Volcanic (slightly weathered)	good	1
Granite–diorite	good	1
Trap	good	1
Magnetite	good	1
Pyrite (disseminated in trap)	good	1
Iron-bearing quartzite	good	1
Sedimentary conglomerate (hard)	good	1
Carbonates (slightly weathered)	fair	3
Carbonates (sandy, medium hard)	fair	3
Sandstone (medium hard)	fair	3
Crystalline carbonates (hard)	fair	3
Crystalline carbonates (slightly weathered)	fair	3
Gneiss (soft)	fair	3
Chert and cherty carbonates	fair	3
Granite (friable)	fair	3
Volcanic (soft)	fair	3
Pyrite (pure)	fair	3
Flints and jaspers	fair	3
Carbonates (soft, slightly shaly)	poor	6
Carbonates (soft, sandy)	poor	6
Carbonates (deeply weathered)	poor	6
Carbonates (shaly clay)	poor	6
Carbonates (ochreous)	poor	6
Chert and cherty carbonates (weathered)	poor	6
Sandstone (soft, friable)	poor	6
Quartzite (fine grained)	poor	6
Crystalline carbonates (very soft, porous)	poor	6
Gneiss (friable)	poor	6
Granite (friable)	poor	6
Encrustations	poor	6
Cementations	poor	6
Schist (soft)	poor	6
Ochre	deleterious	10
Shale	deleterious	10
Clay	deleterious	10
Decomposed volcanics	deleterious	10
Slates	deleterious	10
Talc-gypsum	deleterious	10
Iron formations (very soft)	deleterious	10
Sibley formation	deleterious	10

Sedimentary rocks, because of their variability, may or may not be suitable for aggregate use. The sedimentary rocks of the Gander–Clarenville area that are suitable for aggregate use are massive sandstone, silicious siltstone and graywacke of the Musgravetown and Connecting Point groups provided they are hard, dense, fresh and free of deleterious substances.

Sandstone, siliceous siltstone and graywacke have a petrographic factor range of 1 to 6 with a usual factor of 1 when fresh and hard. These rock types may all be used as crushed stone aggregate for roads, concrete, breakwaters, asphalt, canals, dams, retaining wall construction and railway and road ballast. The sedimentary rocks in the Gander–

Table 2. Revised petrographic factors for some rock types

Rock Type	Petrographic Factor Range	Usual Factor
1. Sandstone	1-6	1
2. Shale	10	10
3. Mudstone	3-6	6
4. Siltstone	1-6	1
5. Conglomerate	1-10	6
6. Arkose	1-6	1
7. Argillite	3-6	6
8. Graywacke	1-6	1
9. Chert	1-3	1
10. Limestone	1-6	1
11. Dolomite	1-6	1
12. Quartzite	1-6	1
13. Granite	1-6	1
14. Gabbro	1-6	1
15. Diorite	1-6	1
16. Granite-diorite series	1-6	1
17. Felsic volcanics	1-6	1
18. Mafic volcanics	1-6	1
19. Intermediate volcanics	1-6	1
20. Felsic-mafic volcanics	1-6	1
21. Pyroclastics	3-6	3
22. Metavolcanics	3-6	3
23. Gneiss	1-6	3
24. Schist	3-10	6
25. Phyllite	6-10	6
26. Marble	1-6	1
27. Slate	10	10
28. Amphibolite	6-10	6
29. Ultramafic	6-10	6
30. Metasediments	1-6	3
31. Iron formation	6-10	6
32. Drift deposits	Any or all of the above	Any or all of the above

Clarenville area that are usually unsuitable for aggregate use are shales and conglomerates of the Davidsville Group and the Brigus and Bonavista formations. Shale, which has a high petrographic factor of 10, because of its softness, friability and mineral content, is not used in most applications as an aggregate. However, it may be used extensively as a raw material in portland cement, as a lightweight aggregate and also as a fill or ballast for road construction. Conglomerate has a high petrographic factor range of between 1 and 10, and a normal factor of 6 when fresh. It is not used normally as an aggregate, because of its weak internal cement, which may break down during handling of the rock. Nevertheless, some cemented homogenous conglomerates (e.g., those consisting of predominantly volcanic or granitic clasts having a silica cement), maybe used as armour stone, for breakwater construction.

Table 3. Affect of weathering on petrographic factors

Petrographic Factor	Weathering Grade	Final Petrographic Factors
1	1, 2	1,2
	3	3,4,5
	4, 5	6,7,8,9
	6	10
3	1, 2	3,4,5
	3, 4	6,7,8,9
	5, 6	10
6	1, 2	6,7,8,9
	3, 4, 5, 6	10
10	1, 2, 3, 4, 5, 6	10

Igneous rocks mainly from the Gander area are subdivided on the basis of coarseness. Medium to coarse-grained rocks (granite, diorite and gabbro) are generally plutonic (intrusive), and the finer grained rocks (rhyolite, andesite, basalt and pyroclasts (tuffs and breccias)) are volcanic (extrusive). These igneous rocks have petrographic factors ranging from 1 to 6 and a normal factor of 1 when fresh.

The medium- to coarse-grained intrusive rocks of the Georges Pond, Middle Brook, Ragged Harbour, Big Round Pond and Warehams granites are all from the Gander area and are all excellent sources of aggregate material, and can be used for both industrial and architectural purposes. The industrial use include crushed−stone road construction, dams, bridges, pavement, retaining walls, railway ballast and foundation blocks. Architectural uses include dimension stone for ornamental and structural use in monumental structures, private, residential, commercial and institutional building complexes. The intrusive rocks that were generally unsuitable for aggregate are from the Newport Granite suite and parts of the Hare Bay Gneiss.

Fine grained extrusive volcanic rocks of the Bull Arm Formation are excellent aggregate materials, but care should be taken when used in concrete. Highly siliceous rocks may react chemically with high-alkali cements, resulting in progressive deterioration of the material. The extrusive rocks of the Love Cove Group are generally not suitable for aggregate material due to their intense alteration and weathering.

Metamorphic rocks may be both suitable or unsuitable for aggregate use depending on their mineralogical characteristics. The metamorphic rocks in the field (Hare Bay Gneiss), which are suitable for aggregate use, consist of granitic-gneiss which are hard and have a normal petrographic factor of 1. The unsuitable metamorphic rocks are slates, pelites, phyllites and schists of the Gander, Mustravetown and Connecting Point groups because of their mineral content, distinct cleavage and softness, and have a petrographic factor of 10 for slate and pelite, 6 for phyllite and schist. These rocks are normally used as fill material.

LABORATORY INVESTIGATION

Petrographic examination (ASTM method C-295.65, 1973) and CSA standard tests A23.2.30 (1973) were carried out on all hand specimens and thin sections made from them. Petrographic examination involves a detailed investigation to determine the quality of the rock sample by identification of such features as microfractures (fractures in the crystal) (Plate 7), porosity in the rock (Plate 8), strained quartz (Plate 9), mineral alteration (Plate 10) and iron-oxide staining in fractures (Plate 11), all of which are deleterious to the rock. Soundness and abrasion tests are being carried out on select samples.

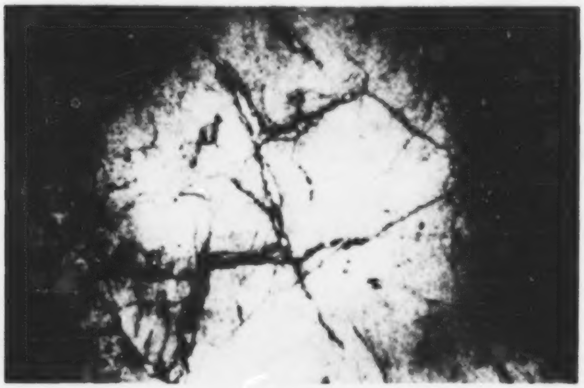

Plate 7. *Microfractures in a quartz crystal; field of view is 5 mm.*

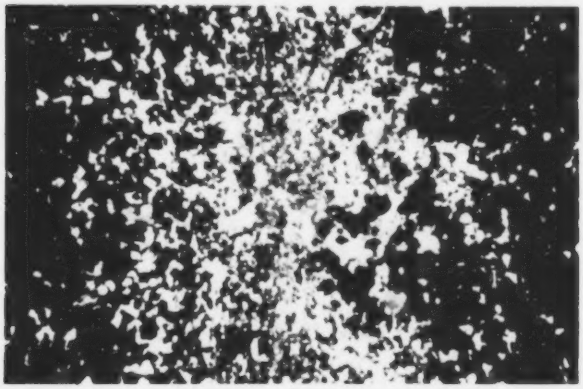

Plate 8. *The darker spots are void spaces that indicate high porosity; field of view is 5 mm.*

SUMMARY

A total of 294 sites were visited and 243 samples were collected during the field season. Thin sections were made of all samples and petrographic analyses is presently continuing.

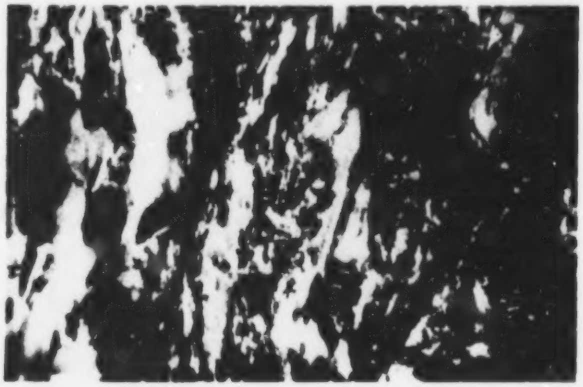

Plate 9. *Strained quartz showing undulating extinction; field of view is 5 mm.*

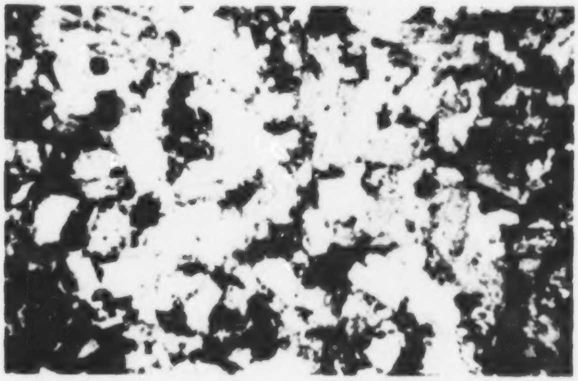

Plate 10. *Feldspars altered to clay minerals; field of view is 12 mm.*

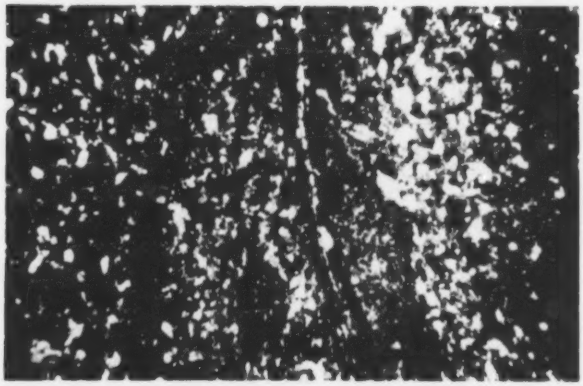

Plate 11. *Iron-oxide staining in fractures; field of view is 12 mm.*

A more comprehensive report is scheduled for publication later in 1989, and will give descriptions of each site and sample. This will eliminate certain sites for aggregate use and promote others. The report will provide a data base for the exploitation of bedrock for aggregate use.

ACKNOWLEDGMENTS

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PLACER GOLD POTENTIAL OFFSHORE NEWFOUNDLAND: A PRELIMINARY ASSESSMENT

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ABSTRACT

A classification scheme and set of sedimentation models have been developed for high latitude marine-placer deposits and are used to assess the relative placer gold potential of the Newfoundland shelf. Temporal and spatial variations in sea level, marine energy and sediment supply are evaluated within eleven geographical zones on the Newfoundland shelf. A comparison is made with the formation controls identified for each deposit type. An estimate of the likelihood of occurrence for each deposit type is provided for each geographical zone. Since there is no basis on which to assess the absolute marine-placer potential of the zones, they are ranked on the basis of their potential, relative to each other.

Of the eleven geographical zones, five are associated with on-land gold mineralization. A closer examination of these indicates that Zone 9, which extends from Twillingate to Indian Bay, has the highest relative marine-placer gold potential. Zone 9 is characterized by a wide range of potential deposit types, including those considered to hold the strongest economic promise.

INTRODUCTION

This paper presents a regional assessment of relative marine-placer gold potential on the Newfoundland continental shelf. The study is part of a larger project that will assess the placer potential of platinum, chromium, iron, zirconium and titanium. A more detailed and comprehensive examination of those sites having the highest potential is in preparation for later publication.

The success of offshore gold placer-mining on the Alaskan continental shelf has resulted in speculation as to the viability of seabed mining in Canadian waters (e.g., Samson, 1984; Hopkins, 1985). West Gold Incorporated is actively mining the Alaskan deposits at a rate of 30,000 ounces per year and has a cash operating cost of less than US $200 per ounce. The total recoverable reserves for the deposit are estimated at 780,000 ounces (Mining Journal, 1986). Tin placers off the Soviet Union and Malaysia have also been the focus of extensive mining activity (e.g., Maslov, 1975; Ringis, 1976; Kudrass and Schulter, 1976), whereas other notable exploration activities include chromite prospects off Oregon (e.g., Kulm, 1988; Petersen *et al.*, 1986; Petersen *et al.*, 1987) and gold prospects offshore of New Zealand (Boyle, 1979; R. Falconer, personal communication, 1988) and British Columbia (Samson, 1984; J.V. Barrie, personal communication, 1988).

Newfoundland has recently experienced a significant growth in mineral exploration. Clearly, where mineralized bedrock and associated overburden sediment are located in proximity to coastal waters, a potential for marine-placer deposits exists. A classification scheme and set of sedimentation models developed for high latitude marine-placer deposits (Emory-Moore and Solomon, *in preparation)* is used here to aid in the regional delineation of such sites on the Newfoundland shelf.

Deposit Classification

Marine placers are subdivided into modern deposits and relict deposits. Modern placer deposits are defined as those that are in equilibrium with the marine processes by which they are formed. Relict placers deposits form under environmental conditions that differ from the modern marine setting (e.g., submerged fluvial).

Modern marine-placer deposits can be classified on the presence and character of a source unit, the marine sedimentary environment in which they form and the mode of mineral concentration (Emory-Moore and Solomon, *in preparation;* Table 1).

1) source: placer deposits that are derived from post-glacial weathering of bedrock are termed 'primary', whereas those derived from the reworking of overburden sediment (i.e., glacigenic sediment) are termed 'secondary'.

This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

Table 1. Deposit classification and assessment factors for high latitude marine placer deposits (cross-hatching indicates preferred choice; from Emory-Moore and Solomon, *in preparation).*

DEPOSIT CLASSIFICATION					CONSTRAINING FACTORS IN THE ASSESSMENT OF HIGH LATITUDE MODERN MARINE PLACERS														
					Sea Level					Wave Energy				Sediment Supply					
					Movement			Rate		Exposure		Shelf Gradient		Source Type				Net Accumulation	
					trans.	stable	regress	fast	slow	high	mod.	steep	shallow	bedrock	offshore	Coastal bluffs	fluvial	high	low
MODERN	Primary Source	beach/nearshore	Autochthonous	1	▨				▨	▨		▨		▨					▨
		offshore		2	▨	▨	▨		▨	▨			▨	▨					▨
	Secondary Source	beach/nearshore	Autochthonous	3	▨				▨	▨		▨				▨			▨
		offshore		4	▨				▨	▨			▨		▨				▨
		beach/nearshore	Allochthonous	5	▨	▨	▨ *		▨		▨		▨		▨	▨	▨	▨	
		offshore		6		▨	▨		▨		▨		▨				▨	▨	
CONSTRAINING FACTORS IN THE PRESERVATION POTENTIAL OF RELICT PLACER DEPOSITS																			
RELICT	fluvial			7	▨			▨			▨		▨						▨
	glacigenic			8	▨			▨			▨	▨							▨
	beach/nearshore			9	▨			▨			▨	▨							▨

* where the source type is fluvial

2) sedimentary environment: the modern marine environment is divided into the beach–nearshore zone and the offshore zone. The beach–nearshore zone is defined as that area shoreward of the breaker zone and extending to the backshore beach complex. The offshore zone extends seaward of the breaker zone to a water depth of 100 m.

3) mode of concentration: offshore-placer deposits can be divided into allochthonous and autochthonous varieties depending on the mode of mineral concentration. Allochthonous deposits can occur hundreds of kilometres from source and generally form discrete lenticular laminae interbedded with well sorted quartzose sands. They form under conditions of net sediment aggradation through a process of hydraulic sorting. Allochthonous placer minerals include zircon, monazite, ilmenite, rutile, magnetite, chromite, and fine grained gold and platinum (i.e., generally < 50 microns). Autochthonous deposits form close to source through grain entrapment within porous gravel sequences or bedrock crevices. They are residual in nature, and occur in areas where the rate of marine erosion exceeds the rate of net sediment accumulation. Autochthonous placer minerals include cassiterite and coarse grained gold and platinum.

Relict deposits are classified on their primary depositional environment only, and include submerged fluvial, glacigenic and beach–nearshore deposits.

Shelf Sedimentation and Marine-Placer Formation

Three primary factors control the processes of shelf sedimentation and, consequently, the formation of marine-placer deposits and the preservation potential of relict deposits. These factors are sea-level fluctuations, sediment supply and the energy of the marine environment (Davis and Clifton, 1987). An examination of the role that each of these factors play in placer formation, and a brief summary of the relevant data for the Newfoundland shelf, is provided below.

Sea Level. As the hydraulic conditions required for offshore-placer formation fall within a critical water depth zone, the net results of eustatic and isostatic variations (e.g., local regressions and trangressions) control the location and extent of a deposit. Relative sea-level movement also influences both the type and volume of sediment supplied to the shelf region. For example, a regressive sea level results in base-level lowering and fluvial incision of sediment; where a river drains into the ocean, large volumes of sediment are supplied. In addition, the rate of sea-level change governs the degree to which sediment is reworked and, thus the likelihood of modern placer formation and relict placer preservation. Where sea-level change is slow, sediment is extensively reworked and the probability of allochthonous and authochthonous placer formation is greatly increased. A rapid sea-level rise favours the preservation of relict placers as sediment reworking and re-distribution is minimal.

The postglacial sea-level history of the Newfoundland shelf is, in most areas, poorly defined (Figure 1). Sea-level curves have been reported for the Great Northern Peninsula (Grant, 1972), western Newfoundland (Grant, *in press)* and southwestern Newfoundland (Brookes *et al.*, 1985). Postglacial sea level along western Newfoundland varies from a major regressive phase in the north to a regressive–transgressive phase in the south. Work on the Grand Banks indicates sea-level submergence of about 120 m (Fader and Miller, 1986) whereas recent work in northeastern Newfoundland indicates a slow regressive–transgressive phase (Shaw and Forbes, 1988).

Sediment Supply. There are four primary sources of sediment and associated particulate gold to the nearshore shelf regions. These are: 1) modern fluvial discharge, 2) erosion of unconsolidated coastal bluffs, 3) erosion of coastal and submarine bedrock faces, and 4) submarine erosion and shoreward transport of unconsolidated sediment. Autochthonous placer deposits form under erosive conditions, where the sediment supply is low, whereas allochthonous placers require a moderate to high sediment supply (see Deposit Classification Section). Relict placers are less prospective where overburden sediment is thick; a low sediment supply is thus favoured.

A significant part of coastal Newfoundland is characterized by steep bedrock faces (Figure 2) and the coastline is generally sediment starved (Forbes, 1984), except in areas where there are large unconsolidated bluffs (e.g., St. George's Bay) (Forbes, 1984) or where fluvial discharge is locally significant. The extent of unconsolidated submarine deposits is poorly known on the Newfoundland shelf and for the purposes of this report, assumed to be ubiquitous. Net accumulation is not considered in this regional study as it is highly localized and not easily measured; it is, however, an important factor in detailed studies.

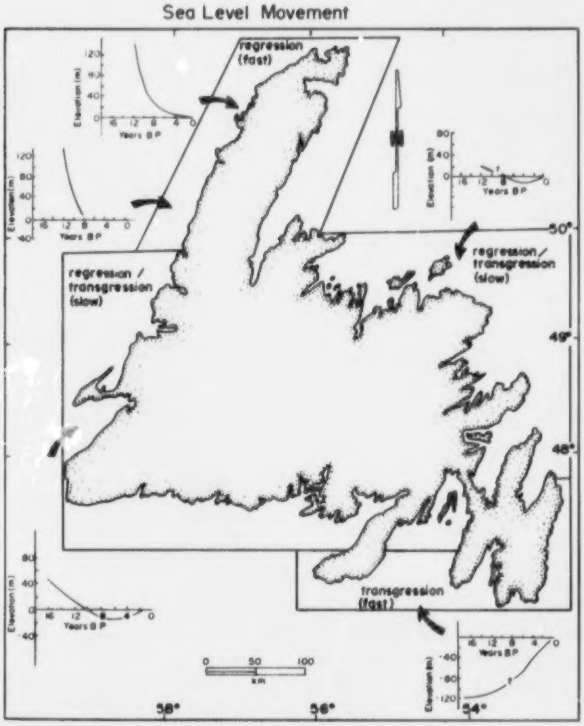
Sea Level Movement
regression (fast)
regression / transgression (slow)
regression/ transgression (slow)
transgression (fast)
Elevation (m)
Years B P
0 50 100
km
50°
49°
48°
58°
56°
54°

Figure 1. *Postglacial sea-level history of the Newfoundland shelf.*

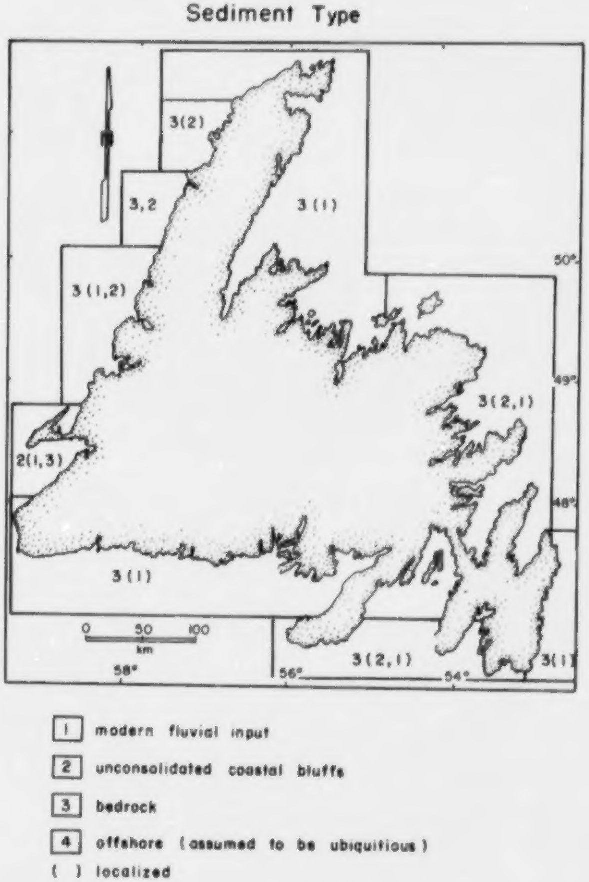
Sediment Type
3(2)
3,2
3(1)
3(1,2)
2(1,3)
3(2,1)
3(1)
3(2,1)
3(1)
0 50 100
km
58°
56°
54°
50°
49°
48°
1 modern fluvial input
2 unconsolidated coastal bluffs
3 bedrock
4 offshore (assumed to be ubiquitious)
( ) localized

Figure 2. *Regional variation in coastal sediment sources.*

Marine Energy. As high latitude coasts are generally storm dominated, relative marine energies can be approximated by evaluating the degree of coastal exposure and to a lesser extent, the shelf gradient. Autochthonous placers form preferentially in areas having a high exposure to oceanic waves, as required for both the erosion of source rock and the erosion and selective removal of light mineral sands. Allochthonous placers form in areas of moderate coastal exposure where sediment is actively reworked and contemporaneously aggraded. Preservation of relict placers is favoured by moderate exposures as the sediment redistribution is minimal.

The gradient of the shelf influences the wave energy and also determines the spatial extent of the zone of offshore-placer enrichment. The hydraulic conditions necessary for the concentration of particulate gold fall within a critical water depth range. On a shallow gradient shelf, this zone will be larger and thus more economically promising than on a narrow shelf. Conversely, a steep shelf favours the formation of autochthonous beach–nearshore placers where very high wave energies are required, and the degree of sediment reworking is not a function of water depth.

The highly irregular morphology of the Newfoundland coastline precludes the definition of any meaningful regional patterns in exposure. Exposure is, however, readily assessed in a detailed study. The gradient of the Newfoundland shelf is highly variable although regional patterns are evident (Figure 3).

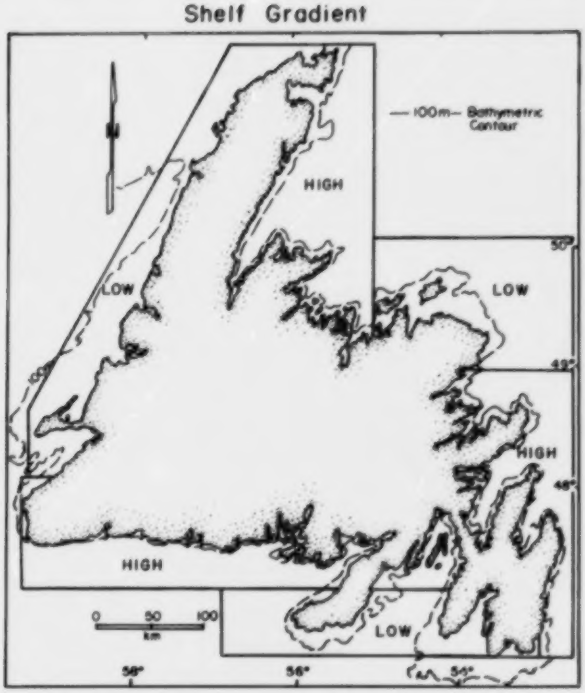

Figure 3. *Regional shelf gradient, offshore Newfoundland.*

Table 1 illustrates the role of the above factors in the formation of individual deposit types. The factors 'source type' and 'sea-level movement' are considered prerequisite factors in placer formation; a source of gold must be identified and mineral enrichment must occur (as controlled by sea level) before a marine-placer deposit will form. The other variables identified above (i.e., the rate of sea-level fluctuations, the net accumulation, the relative exposure and the shelf gradient) influence the effectiveness of enrichment processes and are, thus, not controlling factors but may be broadly considered enhancement factors.

Assessment Procedure

The above classification scheme and formation factors were used to assess the regional modern marine-placer gold potential of the Newfoundland shelf. Shelf areas that are likely to host specific placer types are identified using the following the procedures:

1) review of the literature to establish regional trends in sea-level movement, sediment supply and shelf gradient (Figures 1, 2 and 3);

2) constrain the geographical zonation of the shelf using the previously discussed variables (i.e., sea-level movement, sediment supply, and shelf gradient) (Figure 4); where one factor changes with respect to the other two, a new zone is created. In this manner, eleven zones were established, all having discrete combinations of control factors (Table 2);

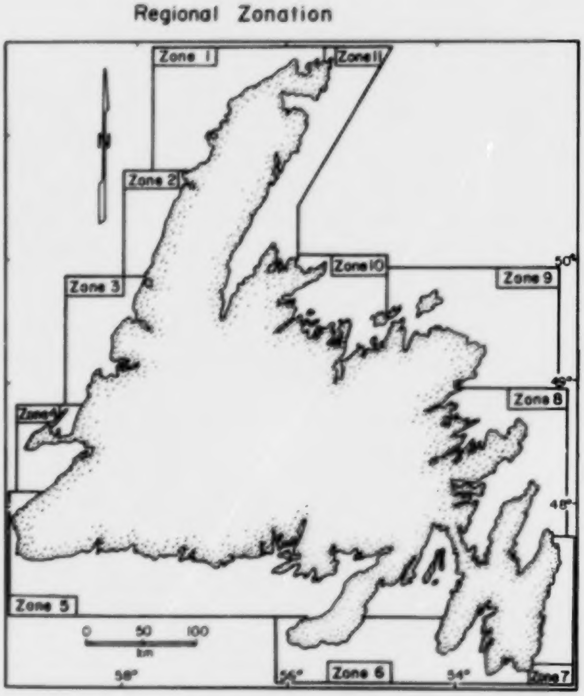

Figure 4. *Zonation scheme used in the regional assessment of marine-placer potential offshore Newfoundland.*

Table 2. Control factors used in the assessment of marine-placer potential on the Newfoundland shelf (dashed check indicates a localized source).

ZONE	Sea Level					Wave Energy				Sediment Supply					
	Movement			Rate		Exposure		Shelf Gradient		Source Type				Net Accumulation	
	trans.	stable	regress	fast	slow	high	mod.	steep	shallow	bedrock	offshore	bluffs	fluvial	high	low
1			✓	✓					✓	✓	✓				
2			✓	✓					✓	✓	✓	✓			
3	✓		✓		✓				✓	✓	✓	✓			
4	✓		✓		✓				✓	✓	✓	✓	✓		
5	✓		✓		✓			✓		✓	✓		✓		
6	✓			✓					✓	✓	✓	✓			
7	✓			✓				✓		✓	✓				
8	✓		✓		✓			✓		✓	✓				
9	✓		✓		✓				✓	✓	✓	✓	✓		
10	✓		✓		✓			✓		✓	✓		✓		
11			✓	✓				✓		✓	✓		✓		

3) compare the prerequisite factors (i.e., sea-level movement and source type) identified for each zone on the Newfoundland shelf (Table 2) with those identified for individual deposit types (Table 1); where both factors are present, the deposit type may occur in the specified geographic zone. The likelihood of occurrence for a given deposit type is then estimated by counting the number of enhancement factors (i.e., rate of sea-level movement and shelf gradient). For example, secondary offshore autochthonous deposits (Table 1, type # 4) are more likely to occur in Zone 9 (Table 2) where the two enhancement factors (i.e., shelf gradient and rate of sea-level movement) are present than in Zone 7, where neither of the enhancement factors are present. The results are presented in Table 3; and

4) those zones where primary gold mineralization occurs in proximity to the Newfoundland coast were identified (Figure 5) and will be the focus of further work.

DISCUSSION

The assessment scheme is restricted to simplified formation conditions and is not amenable to the evaluation of pre-Late Wisconsinan events and their possible role in placer formation. Like classical fluvial placers where multiple rejuvenation and down-cutting events create increasingly rich deposits (Sutherland, 1985), multiple glaciations may have served to alternately concentrate and disperse placers of all types, creating locally enriched glacigenic deposits (e.g., Westland, New Zealand; Henley and Adams, 1979). A further restriction of the assessment scheme is the use of relative indicators (i.e., a 'high' rate of sea-level fluctuation versus a 'low' rate). Information is not available with which to establish absolute values. The assessment scheme does, however, provide a geological framework and rationale for the assessment of relative marine-placer potential.

Table 3. Relative likelihood of occurrence for individual deposit types within the geographic zones on the Newfoundland shelf (brackets indicate localized occurrences).

	DEPOSIT TYPE								
ZONE	1	2	3	4	5	6	7	8	9
1	-	1	-	1	-	-	-	-	-
2	-	1	-	1	-	-	-	-	-
3	1	2	1	2	(2)	-	-	-	-
4	(1)	(2)	1	2	2	2	-	-	-
5	2	1	-	1	1	(1)	-	-	-
6	0	1	(0)	1	1	-	-	-	-
7	1	0	-	0	0	-	-	-	-
8	2	1	-	1	1	-	-	-	-
9	1	2	1	2	2	(2)	-	-	-
10	2	1	-	1	1	(1)	-	-	-
11	1	0	-	0	0	(0)	-	-	-

Of the eleven zones identified on the Newfoundland shelf, five are associated with on-shore primary gold mineralization (zones 5, 8, 9, 10 and 11). Table 3 provides a listing of the deposit types and a ranking of their relative potential for occurrence within each zone.

Given the wide range of deposit types that may occur on the Newfoundland shelf (see Table 3) it is important to establish which deposit types hold the strongest economic promise and where these deposits might occur. The following generalizations may be made:

1) 'primary' placer deposits are limited by the effectiveness of post-glacial weathering and the extent of exposed mineralized bedrock. They are generally low volume although locally high grade (e.g., the beach placers of the Ovens district, Nova Scotia; Samson, 1984);

2) secondary allochthonous placer deposits may attain sufficient grade and tonnage but the particulate gold is frequently very fine grained and, thus, problematic in processing using standard gravity separation techniques; and

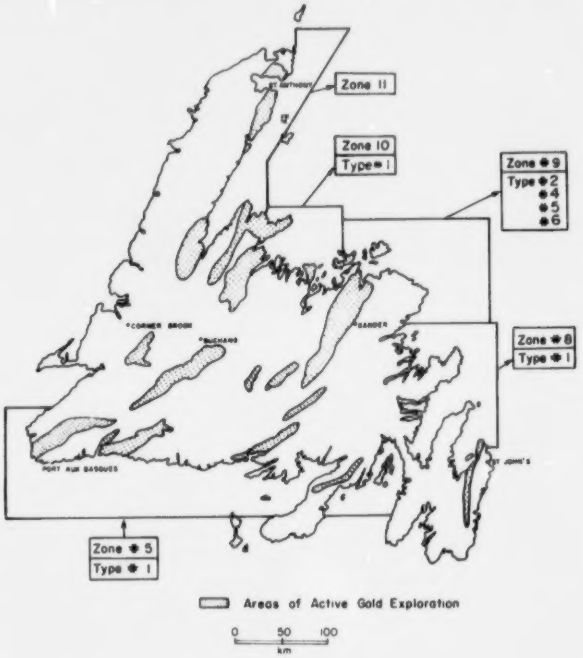

Figure 5. *Regional distribution of gold placer deposit types, which have the highest relative likelihood of occurrence on the Newfoundland shelf (i.e., those deposit types that satisified both enhancement factors; see text).*

3) secondary autochthonous placer deposits are neither restricted by a low availability of source material nor the grain size of gold. In areas of primary gold mineralization, glaciation has resulted in the provision of ample low-grade auriferous material. When subject to marine processes, the extent of the deposit is limited to the critical zone of sediment reworking, which during periods of sea-level transgression, can attain substantial breadth (e.g., gold placers offshore Alaska; Nelson and Hopkins, 1972).

Based on these generalizations, primary placer deposits are considered the least economically prospective. However, primary deposits are easily mined with minimal operating costs. Areas within zones 8, 9, 10 and 11, which are associated with wave erosion of mineralized bedrock, hold a strong potential for this deposit type and are worthy of consideration, particularly by local prospectors. The gold within the mineralized bedrock of Zone 5 is fine grained and, thus, the deposits are restricted to allochthonous varieties.

Allochthonous beach deposits may occur in zones 5, 8, 9 and 10. As the assumed source of sediment in zone 10 is offshore, and given that the sea-level transgression was limited, the potential for allochthonous deposits is less than in zones 5, 8 or 9. In all areas, the formation of allochthonous beach deposits requires a high volume of sediment supply and will be highly localized.

Offshore allochthonous deposits may be found in zones 5, 9, 10 and 11. They are most likely to occur in zones 10 and 11 where a large fall in sea level has generated large volumes of sediment at the mouths of major rivers (e.g., at the head of White Bay).

Areas characterized by a wide transgressive range are the most ideally suited for the formation of 'relict' placer deposits. In this respect, on the Newfoundland shelf, zones 6 and 7 are the most prospective and yet do not host significant gold mineralization on-land. Zones 5 and 8, which are influenced to a lesser extent by a transgressive phase, have the necessary source rock and are thus worthy of closer examination.

Zone 9 has the highest overall marine-placer gold potential with respect to both the wide range of potential deposit types and the high relative potential for secondary autochthonous deposits. A more detailed examination of this area is now underway. The same scheme of assessment will be used but on a detailed scale.

Finally, the assessment of marine-placer gold potential of the Newfoundland shelf as presented here is based on available data, which in some areas is meager. As new data become available, a re-assessment will be necessary.

ACKNOWLEDGMENTS

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QUATERNARY GEOLOGY OF THE BAIE VERTE PENINSULA

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ABSTRACT

Ice flow over the Baie Verte Peninsula was determined by the mapping of ice striations and oriented landforms, and evidence for at least two phases of ice flow were found. The earlier phase consisted of a north to northeast flow over the southern part of the peninsula, diverging east and west over the centre of the peninsula, which acted as an ice divide. The later flow was controlled by local topography, in which the local ice acted independently of the main central Newfoundland ice cap. In the Rambler area, early flow was to the northeast, but later flow shows evidence of an ice divide, with flow both north and south. In an area where striations indicate divergent flow directions, clast fabric analysis was successful in identifying the direction of flow that deposited the surface sediment. Vertical sections in the Indian Brook Valley, suggest the presence of ice-dammed lakes at two separate levels during deglaciation and relict ice wedges indicate periglacial conditions following deglaciation of the valley.

INTRODUCTION

The Baie Verte Peninsula (Figure 1), located in north-central Newfoundland, is of considerable interest with regard to mineral exploration. It has been the site of mines since 1864, having active mining for copper, silver and gold up to 1982 (Hibbard, 1983). In 1988, numerous exploration crews from different mining companies were active on the peninsula, prospecting for gold. Exploration techniques employed by these companies included soil sampling and boulder tracing. The interpretation of results obtained by these methods requires a prerequisite understanding of the ice-flow history, and genesis of surficial sediment. The results of this initial study better define the Quaternary geology of the area, which should help to improve the efficiency and effectiveness of drift exploration on the Baie Verte Peninsula.

Location of the Study Area

The study area is the same as that of Hibbard (1983), bounded by the Trans-Canada Highway to the south, Green Bay to the east, and White Bay to the west (Figure 2). This encompasses all of four 1:50,000 map sheets (2 E/14, 12 H/16, 12 I/1, and 2 L/4) and parts of seven others (2 E/5, 2 E/12, 12 H/7, 12 H/8, 12 H/9, 12 H/10, and 12 H/15).

Bedrock Geology

A comprehensive summary of the bedrock geology is provided by Hibbard (1983). The Baie Verte Peninsula exposes rocks of both the Dunnage and Humber zones, separated by a sharp structural lineament, the Baie Verte line (Williams, 1978). West of the Baie Verte Line are found rocks of the Fleur de Lys belt. These consist of deformed schists and gneisses, intruded by granitoids, and form the eastern margin of the Humber Zone (Hibbard, 1983). Rocks of the Dunnage Zone are found east of the line and consist of ophiolite suites, volcanic cover sequences, and various intrusive rocks, known as the Baie Verte belt (Hibbard, 1983). All the mines and most major showings of economic minerals are associated with the ophiolitic rocks of the Point Rousse and Advocate complexes, and the Paquet Harbour group within the Baie Verte belt, although significant mineralization has been located in the Fleur de Lys belt.

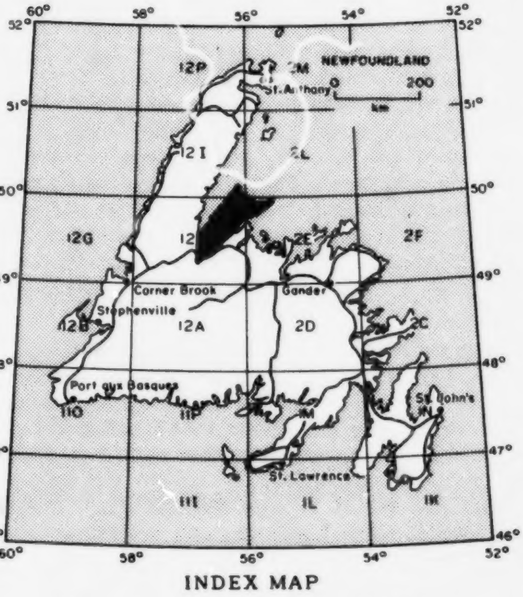

Figure 1. *Index map of the study area.*

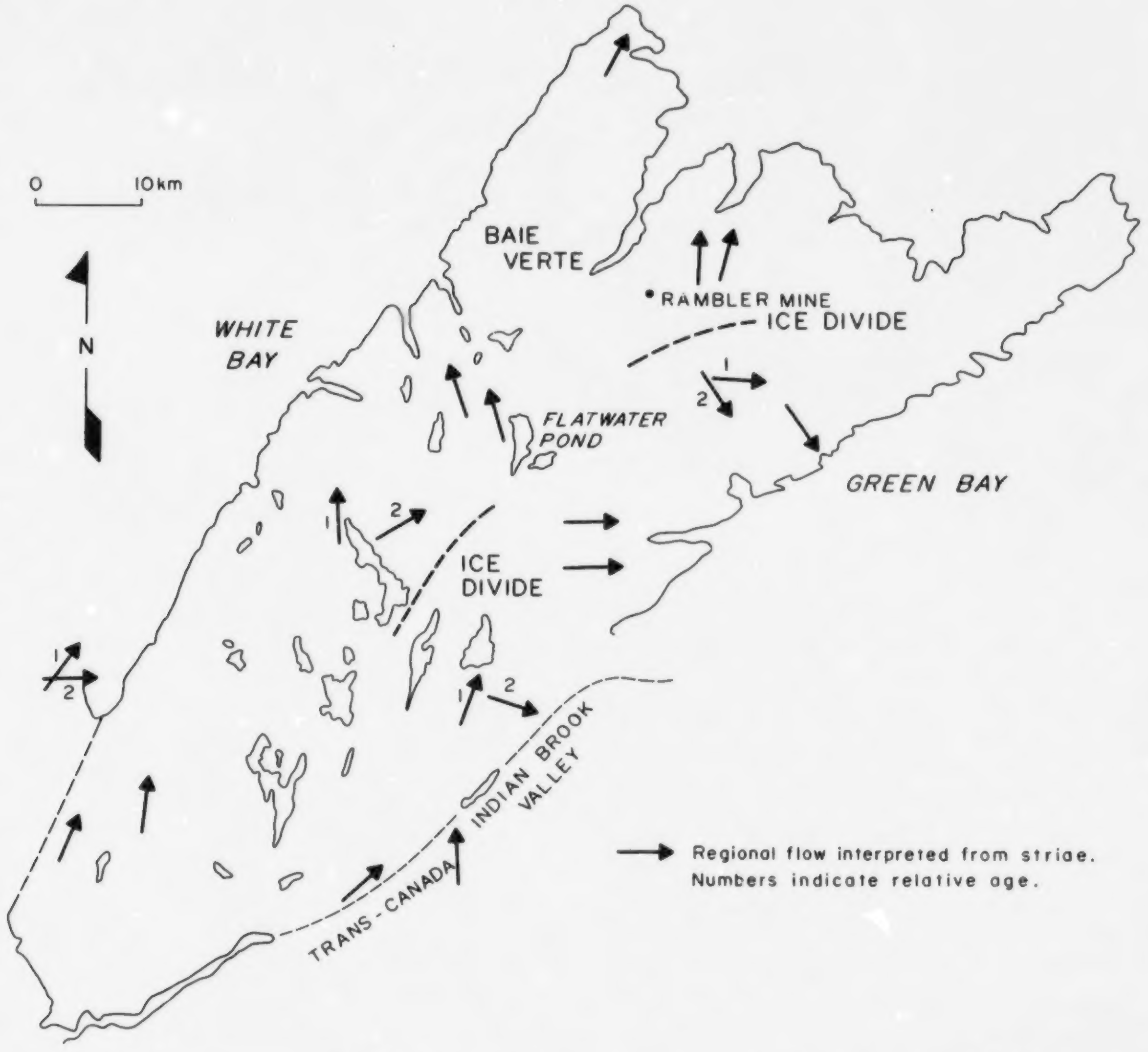
0
10km
N
BAIE VERTE
RAMBLER MINE
ICE DIVIDE
WHITE BAY
FLATWATER POND
GREEN BAY
ICE DIVIDE
1
2
INDIAN BROOK VALLEY
TRANS-CANADA
Regional flow interpreted from striae.
Numbers indicate relative age.

Figure 2. *Ice-flow patterns over the Baie Verte Peninsula. Arrows represent regional flows interpreted from the striation sites.*

Previous Work

Lundqvist (1965) examined ice striations in the area and suggested that there is clear evidence for a single, Late Wisconsinan, major ice flow from the southwest to the northeast over the peninsula. During deglaciation, local topography became increasingly important and caused considerable local variation in flow direction. Tucker (1973, 1974) worked south of the field area, as did Alley and Slatt (1975), and ice striations and oriented-landforms indicate a similar ice-flow history to that determined by Lundqvist (1965).

Grant (1974) suggested that the Baie Verte Peninsula may have supported an independent ice cap during the final stages of Late Wisconsinan deglaciation, with radial flow from the centre of the peninsula. Grant (1986) mapped the peninsula on a reconnaissance level at a scale of 1:50,000, based mainly on aerial photograph interpretation, but with some ground verification.

Vanderveer and Taylor (1987) mapped the Hampden–Sops Arm region on the western margin of the study area, and recognized two regional ice flows; an early west to east flow from ice centred on the Long Range Mountains, followed by a later southwest to northeast flow. Hibbard (1983) provided a map of glacial features, summarized from previous work.

Ice Flow

Figure 2 shows regional ice flows interpreted from over 170 ice-striation sites, including 25 sites from other work

(Lundqvist, 1965; Grant, 1986; Vanderveer and Taylor, 1987), and aerial photographic interpretation of oriented landforms (flutings, drumlins, and crag and tail features), which are not common in the area. Prominent lineaments on aerial photographs commonly parallel bedrock trends rather than reflecting glacial flow. In general, the modern topography appears to be largely bedrock controlled rather than by glaciation.

Ice striations are small-scale erosional features that are formed when clasts transported by ice are moved over bedrock at the base of the ice, and scour the bedrock surface; this results in linear features on a variety of scales. Striations are excellent indicators of ice flow as they are formed by the direct action of moving ice. However, the information from ice striations sites should be treated with caution because of the following reasons:

1) A striation only reflects ice flow at the time of its formation, and should not be used to indicate the total ice-flow history of an area. Earlier or later flows do not necessarily parallel the ice flow recorded by the striation.

2) Striations are relatively small-scale features, and only indicate local flow conditions. Regional ice-flow patterns can show considerable local variation, with ice being deflected by topography and resistant bedrock. Thus regional flow can only be deduced after examining sufficient striation sites to show a regional pattern.

3) The orientation of ice flow can be easily discerned from striations by measuring the azimuth of the linear features. Ice-flow directions can be determined by observation of the striation pattern over the outcrop, where areas in the lee of ice flow may not be striated; by the presence of such features as 'nail-head' striations, and miniature crag and tails; and by the morphology of the bedrock surface, which may show the effects of sculpturing by ice. In many sites, the direction of ice flow is unclear, and only the orientation of ice flow can be deduced.

4) Striations produced by earlier flows can be eroded or obscured by striations produced by later events.

Thus the regional flows interpreted on Figure 2 are all based on at least five striation sites. Where possible, relative age relationships have been interpreted for sites showing more than one striation orientation. These relationships are based on cross-cutting of striation sets, and preservation of older striations in the lee of younger striations. The age relationships plotted on Figure 2 are based on at least three sites.

Ice flow over the peninsula has been complex. The orientation of ice-flow indicators can be explained by postulating two phases of ice flow. The early phase consisted of a north to northeast flow from the central Newfoundland ice cap over the southern part of the peninsula. The centre of the area shows divergence, having eastward flow to the east of the peninsula, and a west to northwestward flow, to the west. This suggests that an ice divide existed in the central part of the peninsula. The later phase consisted of flow controlled by local topography, where local ice acted independently of the central Newfoundland ice cap. In the Rambler area, early-phase flow was to the northeast, but ice flow of the later phase shows an ice divide, with flow to both north and south. During the later phase, ice flowed northeastward along the Birchy Lake–Indian Brook valley. This suggests that ice on the Baie Verte Peninsula was separated from central Newfoundland ice, and may have been flowing south into the valley.

The preliminary interpretations are tentative, as relatively few flow indicators have been located for the size of the study area. Future field work planned for 1989 will check and refine this interpretation.

Relationships Between Ice-Flow Indicators

In mineral exploration, surficial sediment is sampled and analyzed geochemically. Such analyses aim to define geochemical anomalies, related to presence of mineralized bedrock in the area. In glaciated areas, bedrock has been eroded and transported by ice and deposited as surficial sediment. The sediment commonly resulting from deposition in the glacial sedimentary environment is diamicton. Diamicton is a non-genetic term for any poorly sorted sediment showing a range in grain size from clay to gravel, whereas the more familiar term till is restricted to diamictons deposited directly by the action of glacial ice. Thus knowledge of direction of local ice flow is required in locating the source of anomalies.

Striations are oriented parallel to ice flow, but in the case of multiple striation directions, it is not always possible to relate the ice flow deduced from striations to that which moved the sediment being sampled. In glacigenic diamictons, orientation of clasts is often used to infer ice flow, with the clast-long axis oriented parallel to flow (Holmes, 1941; Harrison, 1957). The advantage of this method, known as clast fabric analysis, is that the flow direction obtained moved the sediment being sampled. Clast fabric is also useful in determining diamicton genesis, with strong unimodal fabrics being typical of basal melt-out and lodgment tills, and less well oriented fabrics being found in supraglacial melt-out tills, debris flows, and diamictons produced by ice-rafting of clasts (Lawson, 1979; Domack and Lawson, 1985; Shaw, 1987).

To investigate the usefulness of clast fabric analysis in the Baie Verte area, and the relationship between ice flow, as deduced from clast fabric analysis, and striations, thirteen fabrics were measured from diamicton exposed in trenches in the Flatwater Pond area (Figures 3 and 4). Two striation sites in the vicinity are interpreted as showing an early flow to between 325 and 350°, followed by two subsequent flows at 310 to 320 and 290° (Figure 3). Four other sites show

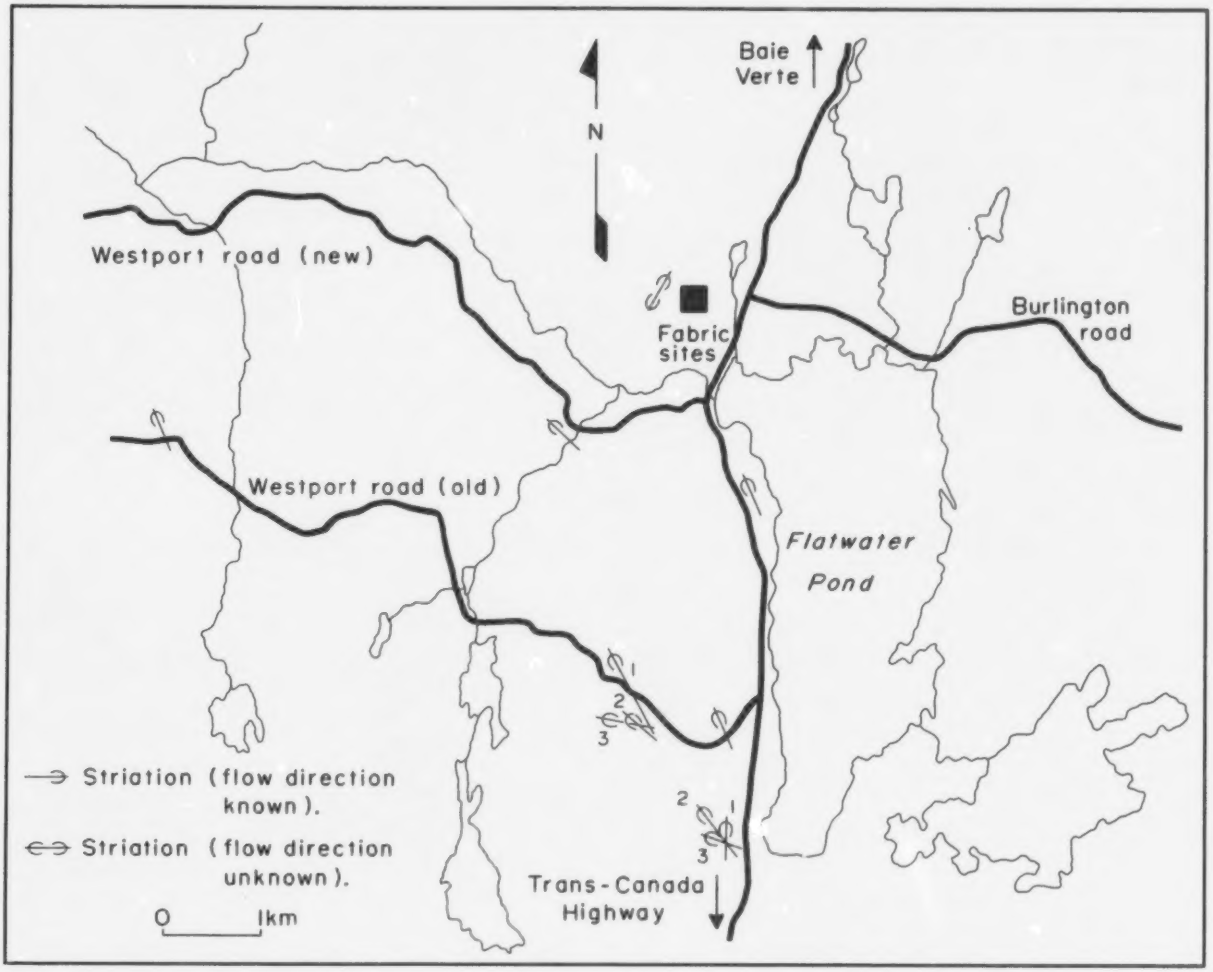

Figure 3. *Striation sites in the vicinity of Flatwater Pond, and location of fabric sites. Numbers indicate age relationships of a set of striation; 3 the oldest, 1 the earliest.*

single sets of striations, oriented between 340 and 350°. A single site in the trenches shows indistinct striations oriented at 037°.

Diamicton exposed in trenches is mainly massive, with a sub-horizontal fissility. Clasts are mainly angular, are of local provenance and constitute between 60 and 80 percent of the diamicton, whereas the sandy matrix forms between 20 to 40 percent. Fragile clasts are 'smeared' along horizontal planes and clasts rarely are aligned along distinct sub-horizontal lines. Resistant clasts are frequently striated. The thickness of the diamicton is variable and controlled by bedrock topography; in lows, diamicton is up to 2.5-m thick, but thins to less than 1 m over bedrock highs.

The results of fabric measurements are variable. Seven sites show strong to very strong unimodal fabrics (S_1 values between 0.72 and 0.92; sites 88-26, 88-28B, 88-28C, 88-28D, 88-61, 88-63B, 88-65; Figure 4). Such fabrics are considered to be typical of basal tills (Lawson, 1979; Dowdeswell and Sharp, 1986; Shaw, 1987). The absence of sorted strata or lenses, dominance of local provenance material, presence of fissility and 'smeared' clasts, suggest that the diamicton is a lodgment till (Muller, 1983; Dowdeswell and Sharp, 1986). Six of these seven sites show a mean orientation suggesting ice flow to between 320 and 356°. Of the sites showing a less strongly oriented fabric, three (88-28A, 88-60, 88-64; Figure 4) show mean orientations between 322 and 002°. Thus it is likely that the dominant sediment-moving ice flow was approximately 330°. This matches the dominant striation direction observed in the surrounding area, and the earliest flow in multiple striation sites.

As deduced from ice-striations, the flow direction suggested is the earliest in the area, and it is possible that the variability in fabrics is due to the re-orientation of clasts by subsequent flow. This would have produced less well oriented fabrics, with a secondary mode to the west of the

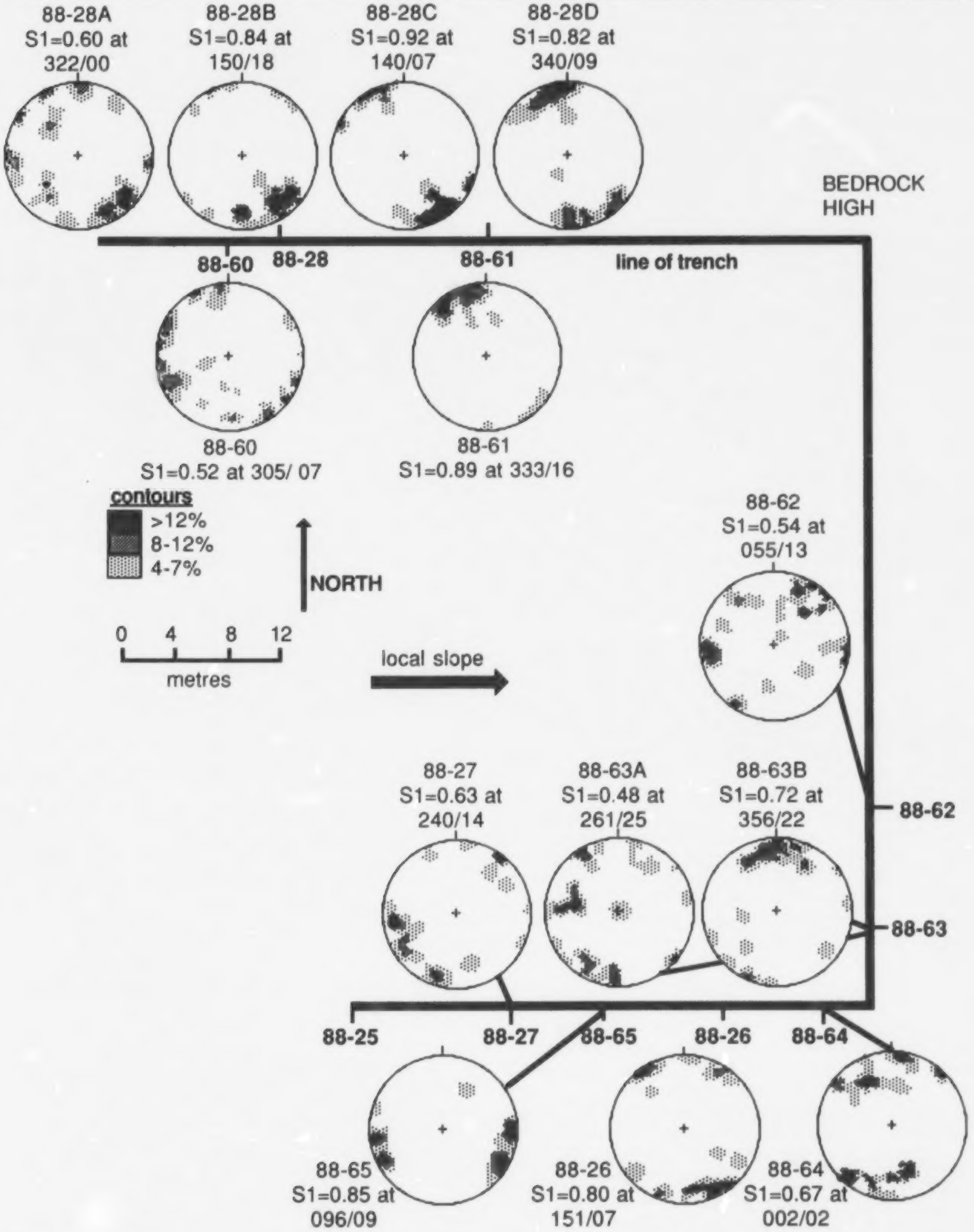
88-28A
S1=0.60 at
322/00
88-28B
S1=0.84 at
150/18
88-28C
S1=0.92 at
140/07
88-28D
S1=0.82 at
340/09
BEDROCK
HIGH
88-60
88-28
88-61
line of trench
88-60
S1=0.52 at 305/ 07
88-61
S1=0.89 at 333/16
contours
>12%
8-12%
4-7%
NORTH
0
4
8
12
metres
local slope
88-62
S1=0.54 at
055/13
88-27
S1=0.63 at
240/14
88-63A
S1=0.48 at
261/25
88-63B
S1=0.72 at
356/22
88-62
88-63
88-25
88-27
88-65
88-26
88-64
88-65
S1=0.85 at
096/09
88-26
S1=0.80 at
151/07
88-64
S1=0.67 at
002/02

Figure 4. *Clast fabrics, Flatwater Pond trenches. The circular plots are equal area stereographic projections of orientations of 25 clast-long axes. They have been contoured using a 2 percent counting circle. The value of S_1 (the principal eigenvalue divided by the sample size) is followed by the azimuth and dip of the mean clast orientation.*

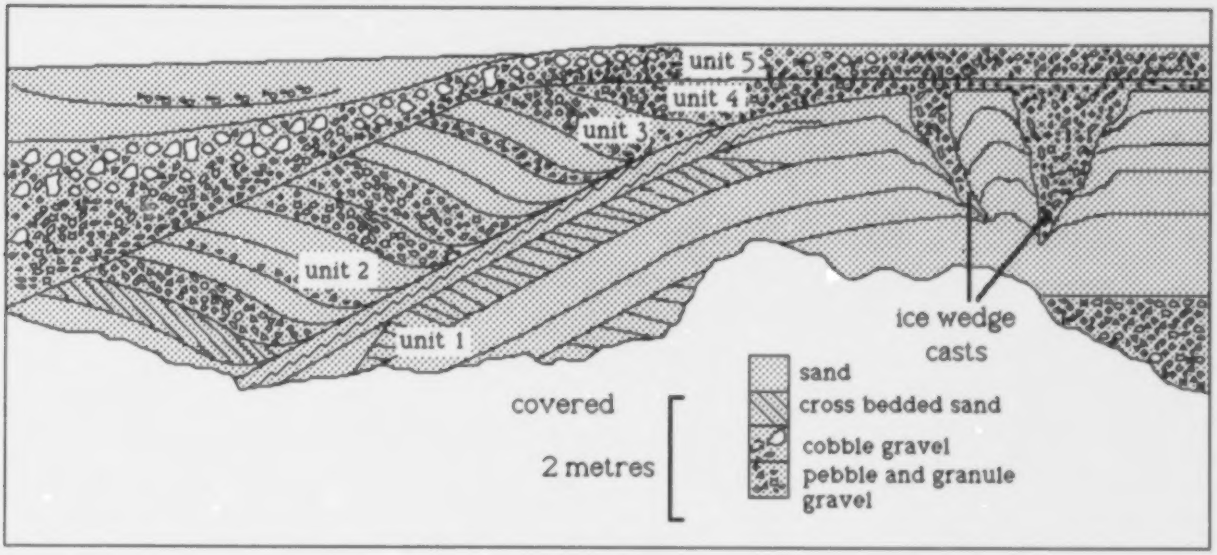

Figure 5. *Sketch of sections in gravel pit at Baie Verte Junction. Unit numbers refer to text.*

primary mode. Dowdeswell and Sharp (1986) found less well oriented fabrics in the upper of two layers in typical lodgment tills. In the cases where two fabrics were taken at different vertical positions at the same site (88-28 and 88-63; Figure 4), the strongest orientations were found at the higher site. Thus it is unlikely that the diversity found is due to either primary depositional processes or re-orientation by later ice flow. Therefore, it is suggested that the main cause of the diversity is post-depositional mass movement. Relief on the bedrock surface is considerable (30 to 40 m), and slopes are steep (up to 20°) and generally dipping to the east. Hence, mass movement of saturated sediment would be likely. The vertical variation in fabric strength and orientation would be due to a series of minor mass movements taking place as cohesive debris flows, with most movement taking place along a zone of dislocation at the base of the sediment. Thus, the upper parts of the diamicton would be rafted relatively undisturbed, but clasts at the base would be re-oriented. Sites 88-27, 88-28A, 88-60, 88-62, and 88-63A (Figure 4) have relatively weak girdle fabrics, which are thought to be typical of debris flows (Nardin *et al.*, 1979; Lawson, 1981).

Evidence from this site suggests that a single fabric measurement may be misleading but multiple fabric measurements are useful in determining sediment-moving ice flow in an area, which had a complex ice-flow history.

Sections in the Indian Brook Valley

The southern boundary of the study area is the valley of Indian Brook (flowing to the northeast), and the Birchy Lake–Sandy Lake system (draining to the west). The drainage divide between the two systems lies at about 140 m above sea level. The valley contains several excellent exposures of Quaternary sediment, and preliminary examination suggests that they may be important in reconstructing the Quaternary history of the Baie Verte area. Two of these sections are discussed in detail.

Baie Verte Junction Gravel Pit. This section is located in a gravel pit lying west of the Baie Verte Highway in the vicinity of the intersection with the Trans-Canada Highway. Between 5 and 8 m of sand and gravel are exposed over 40 to 60 m laterally, forming a cut through the centre of a topographic high (Figure 5). Unit 1, (Figure 5) exposed at the southern end of the pit, consists of 1.5 m of interbedded fine to coarse sand, dipping at 35 to 50° (Figure 5, Plate 1). To the west, the dip decreases over 5 m to close to horizontal. The unit is cut by several small normal faults, which have displacements of 0.5 to 3.0 cm, the fault planes dipping at high angles, and striking approximately north–south. The sediments are mainly either flat stratified, or show planar tabular cross-beds. The cross-beds indicate current flow to the west (in the opposite direction to the primary dip of the beds). This interpretation of current flow is supported by observations of cross-lamination and climbing ripples in fine sand interbeds. Lamination is rarely contorted and folded into tight Z-folds. Sands throughout are moderately to well sorted, and beds are laterally continuous over at least 10 m.

Unit 2 (Figure 5) consists of 2 m of interbedded coarse sand and granule gravel. The beds dip at 5 to 10° to the west, with the dip shallowing up-section. The basal contact of this unit is sharp, truncates underlying beds of Unit 1, and dips to the east at 20°. The relationship of beds with the basal contact is both angular and asymptotic (Plate 1). The unit consists of 80 percent coarse sand and 20 percent gravel, both of which are well sorted. The beds are 10- to 15-cm thick. Gravel is clast supported and the average clast diameter is between 1 and 2 cm, and a maximum of 5 cm. Most beds are massive with some flat bedding and lamination. Planar tabular cross-stratification occurs in one bed. The top of Unit

Plate 1. *South end of sections in Baie Verte junction gravel pit, showing contact between Units 1 and 2; interpreted current flow is to the right.*

2 shows 0.5 m of sand and gravel dipping at a low angle to the northeast. It is overlain by Unit 3 (Figure 5), a 1-m, inversely graded bed of gravel. The base of the unit shows mainly granule-sized clasts, which coarsen upward to cobbles. The gravel is generally poorly to moderately sorted, clast supported, and has a matrix of medium to coarse sand. The largest clasts are about 40 cm long. Unit 3 dips at 10 to 20° to the northeast. It is overlain by Unit 4, consisting of 2 m of interbedded fine to coarse sand and granule gravel, dominated by well-sorted sand. Unit 4 is flat-bedded and cross-bedded having prominent trough and planar tabular cross-beds, suggesting current flow to the west. The upper contact is marked by wedge structures (Figure 5), that penetrate and deform the underlying sand to a depth of at least 1.5 m. On the margins of the wedges, sand beds dip at high angles paralleling the wedge side, and are contorted. Three major wedge structures are seen in this exposure, with the top of wedges being up to 2 m across. The wedges are filled by massive pebble gravels. Many clasts within the wedges are oriented with long axes close to vertical. The gravel of the overlying Unit 5 is moderately sorted, clast supported, and 0.5- to 1.0-m thick, and marks the top of the section.

Interpretation. Examination of the sedimentary structures indicates that the sequence has undergone disturbance. The presence of well-formed, planar tabular cross-beds suggests that the primary bedding surfaces of Unit 1 were originally horizontal. Tracing these beds laterally to the west shows that the dip decreases to near horizontal over 5 to 10 m. The asymptotic contact between Units 1 and 2 suggests that it also was horizontal at the time of sediment deposition. Thus, the overlying beds of Unit 2, originally dipped at steeper angles to the west and may have been deposited as delta foresets. Unit 1 may therefore represent bottomset deltaic deposition, and units 3 and 4, topset beds and fluvial sediments. Deltaic deposition is supported by the well-sorted nature of the basal sediments, with grain-size changes suggesting considerable variation in discharge. Such an interpretation is also supported by the general coarsening-upward trend in the section.

The origin of post-depositional deformation is probably due to the melting of ice buried by the sediment. This style of deformation is unusual, in that the sediment appears to have deformed in a ductile rather than a brittle manner. Dry sand and gravel tend to respond to stresses by failing as coherent blocks, associated with numerous small displacement faults. In this case, only a few minor faults occur, and the displacement on these cannot account for the amount of subsidence seen. It is possible that this plastic-like deformation resulted from the sediment being saturated or frozen.

The wedge structures are interpreted as ice-wedge casts (e.g., Washburn, 1973). Such structures are formed when ice-wedges, formed under permafrost conditions, melt and are infilled by overlying sediment or collapse of the margins. All ice-wedges appear to have formed at the same stratigraphic level, indicating a period of non-deposition, and subaerial exposure ice wedges can form when the mean annual soil temperature is below -5 to -6°C, and there is an adequate supply of moisture (Péwé, 1966). Eyles (1977) reported similar structures in the Botwood area, east of this site.

Thus, this section has a complex depositional history, with indications of variable climate. Current-flow directions throughout the section are to the west, in the opposite direction to the slope and drainage of the present valley. The present altitude of the section is 50 m below the watershed that separates western from eastern flow in the valley. If easterly drainage was dammed by ice, local slope would be toward such an ice dam, and thus water would be impounded, forming a lake, and sediments would have been deposited in a prograding delta. Following drainage of the lake, or synchronous with its existence, ice buried by the sediments melted, which resulted in the collapse and deformation of the overlying material. A period of subaerial exposure followed, allowing the development of ice-wedges. Further sedimentation, probably in a fluvial environment, resulted in the burial and melting of the wedges, and the formation of ice-wedge casts.

Modern permafrost is not recognized in Newfoundland, and Eyles (1977) suggests that ice-wedge formation indicates mean annual air temperatures at least 10°C below modern values. Evidence at the Baie Verte Junction section suggests an initial warm period resulting in deglaciation and formation of an ice-dammed lake, followed by cooling, and a considerable time with lower temperatures to allow the formation of large ice-wedges. Eyles (1977) estimates ice-wedges of similar size would require approximately 2000 years to form. Subsequent burial resulted in the degradation of ice-wedges, and lack of development of periglacial features in upper units of the sequence suggests a warming of climate. No direct chronological control is available for these events. In the Springdale area, there is an extensive complex of raised marine deltas (Tucker, 1973, 1974). Shells from these deltas have provided radiocarbon dates in the range 11,000 to 12,000 BP (Dyck and Fyles, 1963; Lowdon and Blake, 1975; Blake, 1988). Basal lake dates on the peninsula range from 11,800 to 10,400 BP (Blake, 1986, 1987, 1988). It is likely that these marine deltas were deposited, synchronous with or, subsequent to deposition of the sequence described above. It is tentatively suggested that the cold period, resulting in formation of ice-wedges, lay between initial deglaciation of the Indian Brook valley, and the final disappearance of ice from the region.

Upper Indian Brook Sections

In upper Indian Brook, sections extending 300 m laterally, are exposed along a new logging road at approximately 175 m above sea level (Figure 2). Four small sections that represent the entire exposed sequence were chosen for detailed description (Figure 6).

Section 88-81A (Figure 6) contains 2 m of interbedded sand and gravel, dipping at low angles to the south. Gravel constitutes 10 to 15 percent of the section, and is clast supported, granule sized, well sorted, and has a coarse sand matrix. Sand varies from fine to medium grained, associated with a single bed of fine sand to silt. The dominant sedimentary structure is planar bedding, which has some planar tabular cross-stratification that indicates current flow to the south. The fine sand–silt bed is planar laminated at the base, but grades upward to cross-laminated sand. The upper part of this sequence is contorted and folded. The section is capped by 1 m of flat-bedded, clast-supported, pebble gravel. This section is interpreted as showing small-scale delta foresets, with the source of sediment and water to the north of the section.

Section 88-81B (Figure 6) is 4-m thick, with the base lying 1 m above the top of section A and 20 m north of it. The base of the section consists of 1.9 m of interbedded sand and silt and minor gravel. Two units, toward the base, consist of rhythmically laminated fine sand and silt. Each lamina consists of a sharp basal contact, grading normally from fine sand to silt. Laminae are folded and contorted into wide and tight folds. Rare pebbles disrupt the lamination. Overlying these units is 1 m of laminated diamicton. The matrix consists of fine sand and minor clay and silt that constitutes 70 percent of the diamicton. Clasts range from granules to cobbles and deform the lamination underlying them. This sequence is interpreted as showing a transition upward from lacustrine deposition close to a source of sediment at the base, to more distal lacustrine deposition, with input of coarse material by ice-rafting. The sections are exposed on a hillside, and the only possible method of forming a lake in this topographic setting is by damming of drainage by ice.

Section 88-81C (Figure 6) consists of 8 m of diamicton and poorly sorted gravel. The lower part of the sequence consists of diamicton and abundant sorted beds and fine sand to granule gravel laminae (Plate 2). Clast content is 30 percent at the base but increases upward to between 40 and 50 percent. The matrix is mostly sand and some fines, and in places is weakly horizontally laminated. Two fabrics from the basal units show weak to moderate girdle distributions (Figure 6). The genesis of such diamictons is hard to determine, but the weak fabrics combined with the low clast content and abundant sorted strata suggest subaqueous deposition in the proximity of glacial ice. Deposition may have taken place by debris flow, or a combination of suspension settling and ice-rafting. The overlying unit, consisting of a poorly sorted matrix-supported gravel, is very similar to the underlying sediment, but contains few fines. Clasts are mainly pebbles, but cobbles are common, and a few boulders occur.

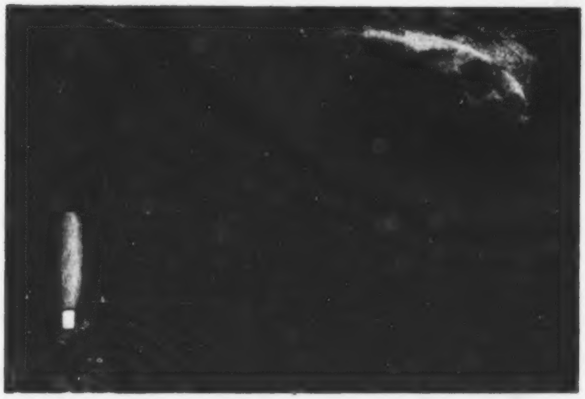

Plate 2. *Structures in diamicton, base of section 88-81C; sorted lense indicates current activity during deposition of the sediment.*

Section 88-81D (Figure 6) consists largely of clast-supported, poorly sorted gravel and interbeds of well-sorted fine to coarse sands (Plate 3). The gravel is massive or crudely stratified and contains striated and faceted clasts of mainly pebble size. Some cobbles and boulders are present. The matrix is a fine to coarse sand associated with rare, well-sorted laminae of fine to medium sand. Interbeds of sand are well sorted, and are mainly massive or planar stratified.

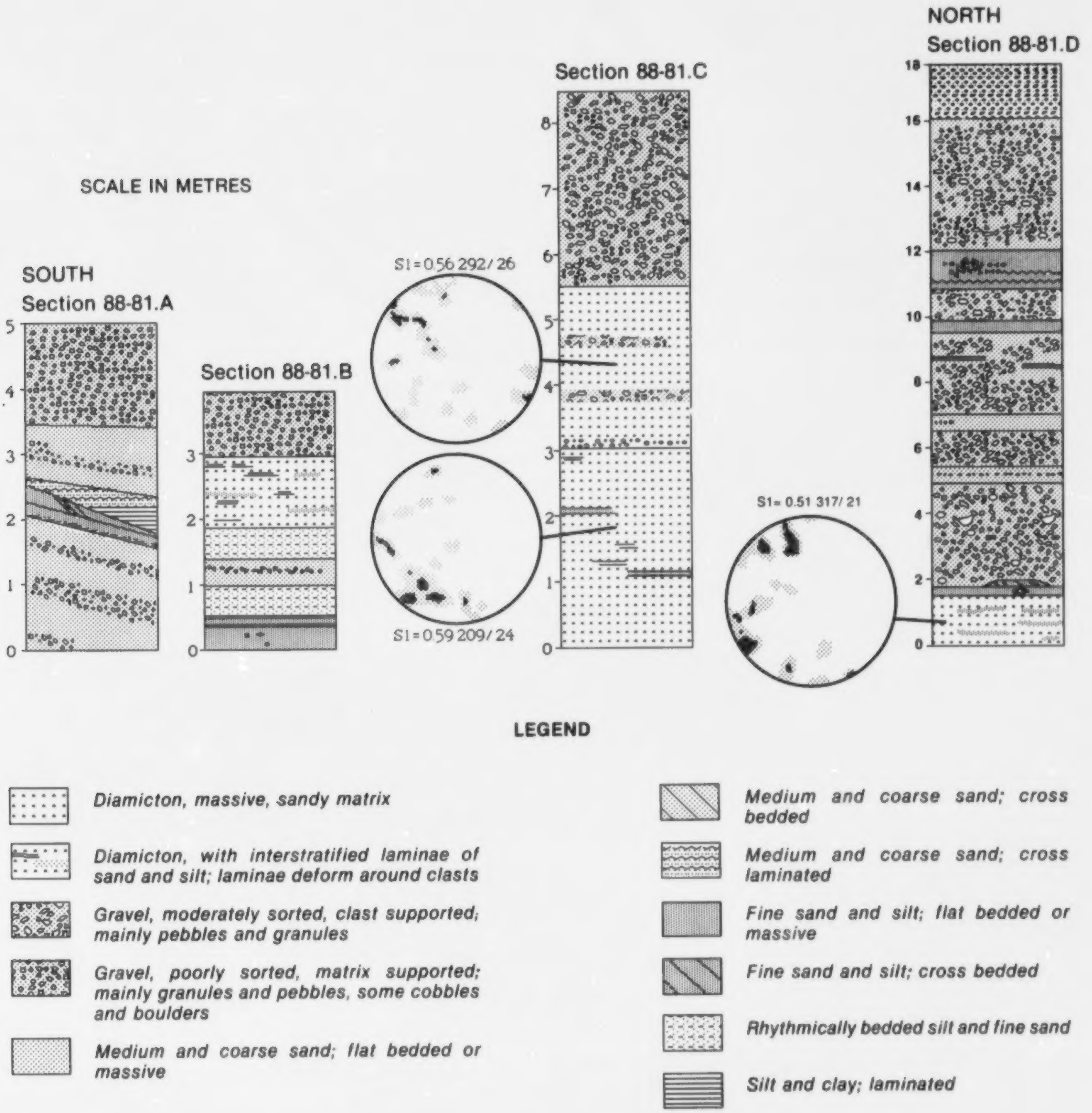
NORTH
Section 88-81.D
Section 88-81.C
SCALE IN METRES
SOUTH
Section 88-81.A
Section 88-81.B
S1=0.56 292/26
S1=0.59 209/24
S1= 0.51 317/21
0
1
2
3
4
5
6
7
8
10
12
14
16
18
LEGEND
Diamicton, massive, sandy matrix
Diamicton, with interstratified laminae of sand and silt; laminae deform around clasts
Gravel, moderately sorted, clast supported; mainly pebbles and granules
Gravel, poorly sorted, matrix supported; mainly granules and pebbles, some cobbles and boulders
Medium and coarse sand; flat bedded or massive
Medium and coarse sand; cross bedded
Medium and coarse sand; cross laminated
Fine sand and silt; flat bedded or massive
Fine sand and silt; cross bedded
Rhythmically bedded silt and fine sand
Silt and clay; laminated

Figure 6. *Stratigraphic columns (sections 88-81A to 88-81D) exposed in the Upper Indian Brook area.*

Plate 3. *Laminated sand and silt in section 88-81D; poorly sorted gravel is probably ice-rafted and either dumped or formed when ice grounded, and melted in place.*

Toward the top of the section, a thick sand interbed is cross-laminated, indicating current flow to the west. The section is capped by 2 m of moderate, to well-sorted, matrix-supported gravel.

Matrix-supported, poorly sorted gravel is probably formed either by mass movement, or by ice-rafting and rain-out of clasts into an environment where current action winnows fines from the substrate (Nardin *et al.*, 1979; Eyles and Miall, 1984). In this case, probably both processes were operating, with the sand interbeds indicating a moderate energy, possibly pro-delta, environment.

In summary, the sections in upper Indian Brook valley contain relatively thick sequences that might be interpreted as till in poor exposures. The sedimentary characteristics of the diamictons in association with probable deltaic and lacustrine sand and silt suggest that the sequence represents deposition into a glacial lake. The diamictons and poorly sorted, matrix-supported gravel units were probably deposited by a combination of mass movement and ice-rafting. This implies that ice was adjacent to, or moving into, the lake.

The elevation of the sections is between 160 and 180 m (estimated from topographic maps). It is possible that the lake was an ice marginal feature, formed when ice filled the Indian Brook valley and dammed southern drainage from the Baie Verte Peninsula. Such a lake would be a minor feature. Alternatively, the lake thought to exist here may be related to the lake postulated from evidence at the Baie Verte Junction section. If so, the lake was large and may have been connected to the Sandy Lake—Deer Lake basin, to the west. To produce thick diamicton sequences in this setting, ice may have been moving south from the Baie Verte Peninsula into an ice-dammed lake in the Indian Brook valley.

Implications for Drift Exploration

The preliminary ice-flow history described here is very complex having considerable local variation. Any attempt to relate anomalies in surficial sediment to bedrock sources should take this into consideration. In problematic areas, careful mapping of ice striations may assist in defining ice-flow history. Clast fabric analysis may be useful in defining sediment moving ice flow in areas of complex or divergent flow, but multiple measurements are recommended. Drift sampling in areas adjacent to the Indian Brook valley may encounter thick sequences of diamicton, deposited subaqueously. Geochemical results from such sediment may be misleading or difficult to interpret, as transport and deposition in ice-proximal glaciolacustrine sequences tend to be very complex.

CONCLUSIONS

The major results of 1988 field work are:

1) The ice-flow history for the Baie Verte Peninsula is complex, with at least two major phases of flow, and with considerable diversity of flow direction;
2) ice divides existed in the Flatwater Pond region, and in the area of Gull Pond (south of the La Scie road);
3) the striation database is small considering the size of the study area and in planning and interpreting the results of geochemical data, consideration should be given to examination of local flow patterns by more detailed mapping;
4) clast fabric analysis may be a valuable supplement to striation and landform mapping in areas of complex ice flow; such work is only useful if the sedimentology and genesis of the till being sampled is well understood;
5) sections in the Indian Brook valley suggest a complex history of deglaciation with evidence for glacial lakes existing at two elevations; and
6) observations of relict ice-wedges confirm the earlier conclusion of Eyles (1977) that severe periglacial climatic conditions followed deglaciation in north-central Newfoundland.

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QUATERNARY GEOLOGY OF SOUTHWESTERN NEWFOUNDLAND

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Terrain Sciences Section

ABSTRACT

A mapping survey was undertaken during the 1985 and 1986 field seasons in the La Poile (11O/9), La Poile River (11O/16), Peter Snout (11P/13) and King George IV Lake (12A/4) map areas. Eleven hundred and thirty-six samples were collected from shallow hand-dug pits, from the 'C' soil horizon in lodgement and melt-out till units. A representative pebble fraction was obtained at 450 sites and was matched to the bedrock map units. The Quaternary sediments have been divided into six units: these are: 1) colluvium, 2) alluvium, 3) glaciofluvial, 4) hummocky deposits, 5) till veneer, and 6) streamlined forms. The 63 micron fraction of the samples were analyzed for Cu, Pb, Zn, Co, Ni, Mn, Fe, V and Au. The 8 mm and finer fraction were analyzed for particle-size distribution. Ice flow in the area was predominantly to the south and extended to at least the coastline.

An analysis of till composition (pebble fraction) was done to determine the till dispersion. Analysis of some of the data for the fluted forms, suggest that about 60 percent of the clasts are locally derived. Pearson correlations between particle size and geochemistry indicate that Pb, Zn and Mo values are influenced by the matrix texture.

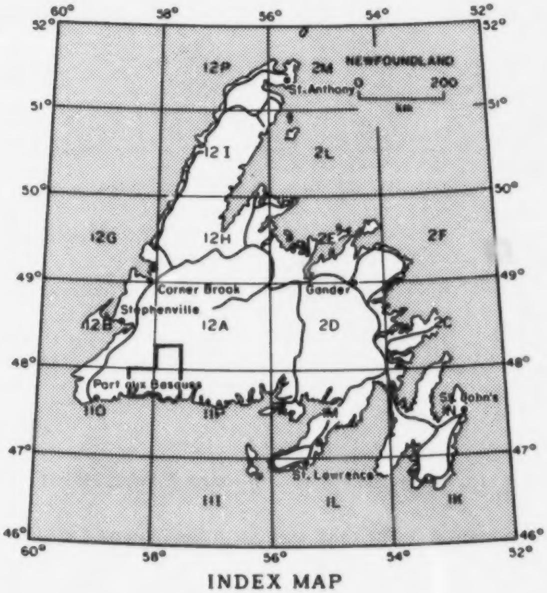

Figure 1. *Location of study area.*

INTRODUCTION

During the 1985 and 1986 field seasons, mapping of the Quaternary geology in the La Poile (11O/9), La Poile River (11O/16), Peter Snout (11P/13) and King George IV (12A/4) map areas (Figure 1) was undertaken. This project is a continuation of a provincially funded Quaternary mapping program, designed to provide information on the glacial dispersion, the distribution of glacigenic sediments and the geochemical and textural characteristics of till, as an aid to exploration in areas of mineral potential. There are no previously published reports on the Quaternary geology of southwestern Newfoundland, and the only map with any information is the Peter Snout area (11P/13) (Grant, 1974).

PHYSIOGRAPHY

King George IV Lake (12A/4)

Most of the King George IV Lake map area is a barren, hummocky upland at an elevation of 350 to 400 m, however, higher hills stand above this level, at a maximum height of 552 m. Three major river systems, the Lloyds, the Victoria and the Burnt Pond rivers drain toward the northeast. The topographic morphology and main drainage pattern reflect the bedrock structures, although many of the smaller lakes and ponds show no discernible geological control. Some ponds, particularly in the area east of Rocky Ridge Pond (Figure 2), appear to reflect southward ice flows.

Peter Snout (11P/13)

The Peter Snout map area is characterized by large areas of barren plateau at an average elevation of about 350 m. Numerous peaks over 500 m constitute part of the Blue Hills of Couteau, which are located to the northwest. Drainage is toward the south via the Grandy Brook system, and an unnamed brook that flows into Connoire Bay, west of Burgeo. The drainage pattern is transverse to the regional geological structure and therefore may be controlled by cross-faults.

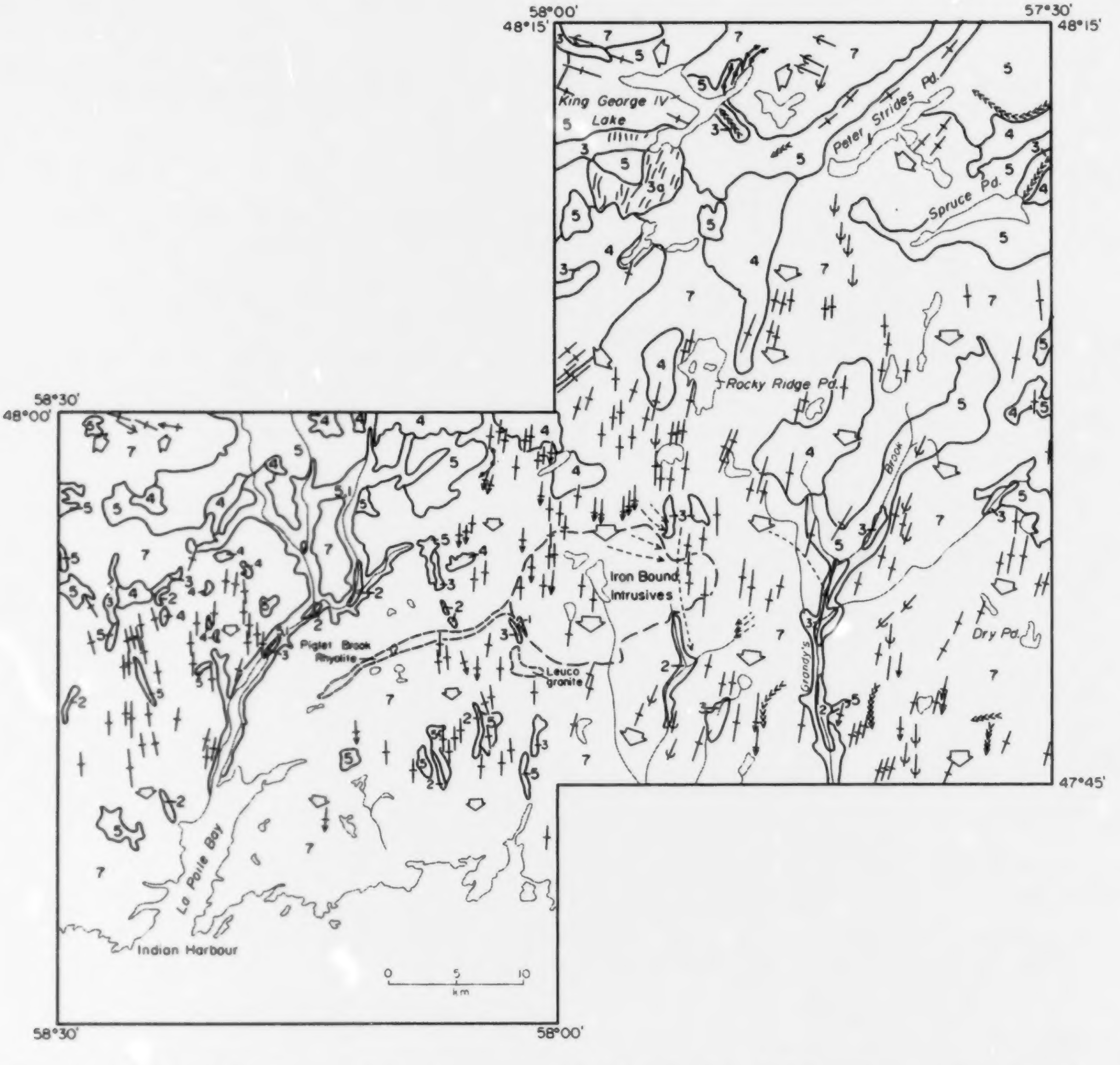
58°00'
57°30'
48°15'
48°15'
King George IV Lake
Peter Strides Pd.
Spruce Pd.
Rocky Ridge Pd.
Brook
58°30'
48°00'
Iron Bound Intrusives
Pigeon Brook Rhyolite
Leuco granite
Grandy's
Dry Pd.
47°45'
La Poile Bay
Indian Harbour
0
5
10
km
58°30'
58°00'

Figure 2. *Generalized Quaternary geology of map areas 11P/3, 12A/4, 11O/9 and 11O/6, southwestern Newfoundland.*

La Poile and La Poile River (11O/9–11O/16)

In this area, the topography is controlled, largely, by the relative resistance of bedrock type and the east-northeast-trending structural grain. The area west of La Poile River and Bunker Hill Brook is characterized by a gently rolling upland underlain mainly by the Rose Blanche Granite. This upland area has an elevation ranging from 520 m in the north to about 200 m, west of North Bay. East of La Poile River, the terrain is rugged and irregular, and the elevation ranges from 450 m in the northeast to about 250 m in the south. This area is underlain by metamorphosed felsic volcanic and sedimentary rocks, amphibolite, gabbro and granitoid intrusions. The Iron Bound intrusives (granite, diorite) at the eastern margin of the map area form the highest hills (593 m) and are a prominent landmark.

GEOLOGICAL SETTING

The King George IV Lake (12A/4) and Peter Snout (11P/13) map areas lie within the southern end of the Paleozoic Central Mobile Belt (Kean, 1983). Both areas contain elements of the Dunnage Zone, and the Peter Snout area also has elements of the Gander Zone. The oldest units consist of ophiolite rocks overlain by an island-arc sequence. These are unconformably overlain by fluviatile sedimentary and subaerial volcanic rocks that have a northeast–southwest trend in the northern part of the area and an east–west trend

LEGEND

QUATERNARY

1 Colluvium deposits: *usually angular material of any particle size, accumulated on the lower parts or the base of slopes and transported by gravity through cliff failure and mass wasting*

Non-glacial environment

2 Sand to bouldery gravel alluvium: *up to 2 m thick, contains terraces and plains associated with stream channels and floodplains; commonly found in the larger valleys and may overlie glaciofluvial sediments*

Pro-glacial environment

3 Ice-contact or glaciofluvial deposits: *consists of gravel and sand in the form of hummocks, terraces and plains; commonly within valleys;* 3a, ribbed moraine: *bouldery diamicton with a sandy matrix, occurs as straight to sinuous ridges 2 to 25 m high and up to 1.5 km long, oriented transverse to flow; occur in valleys and areas of low relief.*

Glacial environment

4 Hummocky moraine (1 to 10 m thick): *commonly consisting of sandy till, but may be composed entirely of boulders, (appears to be of a complex origin, possibly the result of ice disintegration)*

5 Featureless till veneer: *commonly discontinuous and less than 2 m thick; in some areas it consists entirely of boulders and includes colluvium deposits in deeply incised valleys*

6 Linear forms consisting of silt–clay till; *commonly overlain by a thin (1 m) mantle of a more sandy melt-out till containing more boulders; long, flat-profiled flutes are the most common form–other forms include till ramps and ramped crag-and-tail hills*

PRE-QUATERNARY

7 *Bare rock exposed or obscured by vegetation, includes areas of boulder-strewn bedrock*

SYMBOLS

⇨ *Direction of regional ice flow*

⇗ ⤢ *Striation (ice direction known, unknown)*

◌ *Cirque*

Longitudinal ice-flow features

—+— *Drumlinoid, fluting*

+——> *Crag-and-tail hill*

——+> *Ramped hills*

—+—> *Ramped crag-and-tail hills*

Transverse ice-flow features

•-•-• *Crestline of moraine*

in the south. Intrusive rocks range from Silurian to Carboniferous; the older intrusives are concordant with structural trends and the younger ones cross-cut them. The faults trend northeast–southwest in the north and east–west in the south. The rocks vary from undeformed to polydeformed and the metamorphic grade varies from sub-greenschist facies to amphibolite (Kean, 1983; O'Brien, 1983).

The La Poile and La Poile River map areas are mainly underlain by three metavolcanic and metasedimentary units; the Bay du Nord Group, the La Poile Group, hosting the Hope Brook gold deposit, and the Keepings gneiss. These units are intruded by several generations of granitoid rocks, the most prominent being the Rose Blanche Granite, the La Poile batholith and the Chetwynd Granite. All units were affected by east-northeast-trending fault systems, such as the Cape Ray and Bay D'est faults (Chorlton, 1978, 1980).

FIELD PROGRAM

Eleven hundred and thirty-six samples were collected from shallow (50 to 75 cm) hand-dug pits, on an approximate 2- by 2-km grid spacing. Ninty-five percent of these samples are from the *'C'* soil horizon; the remainder are from either the *'B'* horizon or the *'B–C'* transitional zone, and were collected from drumlins and drumlinoid forms, crag and tail hills, ramps, hummocks, eskers and in areas covered by a thin veneer of featureless glacial sediment (commonly till). Where possible, a representative pebble fraction (16 to 60 mm) was obtained at approximately 450 sites. The pebble types were determined and referenced to a bedrock map unit. A landform classification map has been completed for all map areas (Sparkes, 1987) using 1:50,000 scale, black and white airphotos. This interpretation was extensively ground verified during the course of field work, and was used for determining ice-flow directions, possible ice margins, and the distribution of surficial sediment.

SEDIMENT ANALYSES

Geochemical

The 63 micron fraction (-250 mesh) of the matrix samples have been analyzed for Cu, Pb, Zn, Co, Ni, Mn and Fe by atomic-absorption spectroscopy, Au by fire assay and delayed counting neutron activation and U by neutron activation. The finer than 63 micron fraction was chosen for geochemical analyses because of the relative ease of sample preparation, in comparison to obtaining the clay fraction by centrifuging.

Particle-Size Distribution

The 8 mm and finer fraction of the samples have been analyzed for particle distribution from 8 mm to 0.00126 mm. The procedure involves both mechanical sieving through a bank of sieves, and an electronic size analysis using a fine particle analyzer (Coulter counter).

Glacial Flow

Ice flow in the area was predominantly to the south (180 to 210°), as indicated by numerous north–south oriented depositional forms such as flutes, drumlins, crag and tail hills, and till ramps. Glacial striae are not well preserved in the study area, but many bedrock stoss and lee forms indicate ice flow, parallel to the linear features. A diversion of southward-flowing ice in response to local topography is apparent in the vicinity of La Poile Bay and Grandy's Brook (Figure 2). This change is evidenced by a slight reorientation of the streamlined forms and it may also indicate a thinning of the ice near its margin.

In the northeastern part of the King George IV Lake map area, ice flow is more varied (Figure 2), having moved: a) to the southwest, parallel to Peter Strides Pond, b) south in the central part of the area between Spruce Pond and Rocky Ridge Pond, and c) west in the area of King George IV Lake. These divergent flows may be attributed to relatively thin ice conditions, which resulted in flow being controlled by topography and the direction of regional slope. The influence of regional slope upon ice flow is also evident in the northwestern corner of the study area, where linear forms indicate a northwest flow of ice into St. George's Bay.

REGIONAL QUATERNARY GEOLOGY

Figure 2 summarizes the distribution of glacial sediments within the field area. Units 1 and 2 consist of colluvium and alluvium respectively, and were not extensively sampled or studied in any detail. They are of very limited extent, and their occurrences were determined almost solely from airphoto interpretation. Unit 3 is a glaciofluvial sediment, consisting of sand and gravel associated with outwash systems, and is found within most valleys and broad linear depressions. Unit 3a consists of moraines, oriented transverse to glacial flow, which occur as straight to sinuous ridges 2- to 25-m high and up to 1.5-km long. These were formed either at or near the margin of the retreating ice.

The northern part of the study area, and in particular the area west of Rocky Ridge Pond, is characterized by widespread low-relief hummocky till deposits (Unit 4). These hummocks consist mainly of coarse till, which have many large boulders of predominantly local origin, that probably owe their present emplacement to both direct subglacial meltout and supraglacial meltout from stagnant ice. Basal deposition of material from a boulder-rich debris zone (due to stagnant ice conditions) can produce a low-relief hummocky terrain, since the debris load within this zone would not be uniform in its distribution.

Areas of featureless till veneer (Unit 5) occur throughout the study area, but are mainly found in the north. This unit is discontinuous, less than 3-m thick and commonly consists of a locally derived boulder lag. The form of this unit mimics the morphology of the underlying bedrock, and the mode of deposition may be similar to that of hummocks, i.e., meltout from stagnant ice.

The most widespread depositional features are linear forms that are parallel to regional ice flow (Unit 6, see symbols for Figure 2), and generally occur in areas lacking other sediment types. These features can be subdivided into three major types: flutes, till ramps, and ramped crag and tail hills. The fluted forms are up to 2-km long, 5- to 15-m high, have a relatively flat profile, and are most common in areas of low relief. Till ramps occur up-ice of bedrock knobs and are up to 1.5-km long, 400-m wide and 30-m high. Several of the ramped hills also have tails that are up to 1.3-km long, 200-m wide and 30-m high.

Many mechanisms of flute formation have been proposed (Galloway, 1956; Boulton, 1971; Smally and Unwin, 1968; Shaw and Freschauf, 1973; Jones, 1982) and all involve complex interactions between erosional and depositional processes. It is generally believed that these features develop in a subglacial environment and require an obstruction in the glacier's path to initiate deposition. This may be an embedded boulder, a bedrock knob, or a lodged block of frozen basal debris. Deposition of material down-ice of such an obstruction may be accomplished by lateral particle migration (plastic deformation) into a low-pressure zone within the lee-side cavity of the obstruction, or by subglacial meltout, if thawed-bed conditions occur. Lateral particle migration may explain the lack of any other sediment type in areas where the linear forms are most abundant and closely spaced, since the growth of the fluted forms would be at the expense of deposition in the intervening areas. Alternatively, these areas may be devoid of sediment due to net erosion.

Till ramps, unlike other streamlined forms, develop on the up-ice side of bedrock obstructions and are a common depositional feature in this part of the southwest coast. Boulton (1971) observed till deposition on both the up-ice and down-ice sides of a bedrock knob, underneath the margin of a glacier in Svalbard, Spitsbergen. He suggested that the up-ice deposition is partially the result of pressure melting of the glacier sole and the subsequent gradual accumulation of particles released from the ice. The deposition of materials down-ice of the knob (ramped crag and tail hills) was in a lee-side cavity, and occurred by squeezing and slumping of subglacial till into the cavity. Although this type of deposition was observed by Boulton (1971) on a relatively small scale (tens of metres), it may be applicable to these large-scale forms in the study area. The material in these ramps consists of a very compact, possibly lodgement till, exhibiting preferred pebble orientations parallel to the regional ice flow (Figure 3). These characteristics are consistent with those observed by Boulton (1971) for the subglacial material found up-ice of bedrock knob beneath the margin of modern glaciers. The absence of material down-ice of these forms in southwestern Newfoundland, may be the result of a relatively small sediment load and, therefore, not enough basal material is provided for much of any deposition by squeezing or slumping into a lee-side cavity.

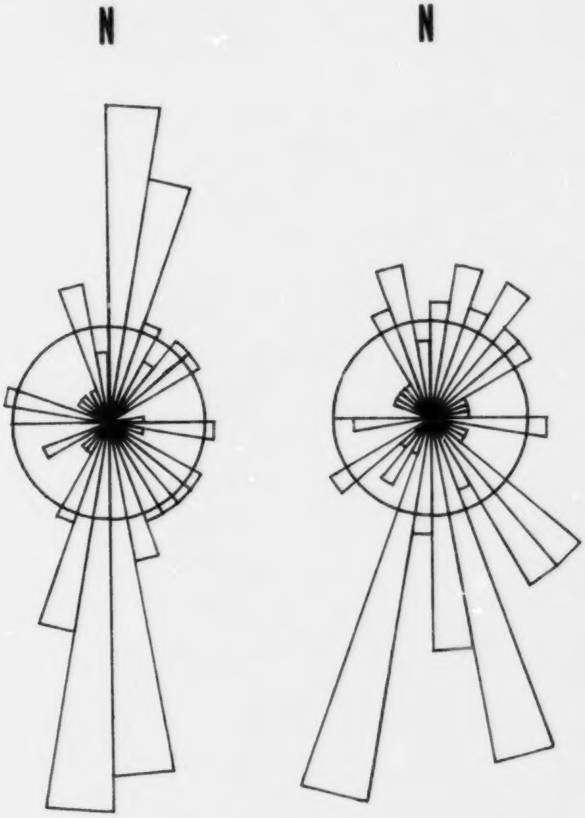

Figure 3. *Rose histograms for clast orientations on a typical till ramp.*

Glacial Dispersion

An analysis of till composition was undertaken as part of the study. A representative pebble fraction (16 to 64 mm) of the till was obtained at 300 sites, mostly from streamlined forms. Approximately 75 clasts were collected and identified from each hand-dug pit (75- to 100-cm deep), and where possible, the clasts were related to their bedrock source.

Analysis of some of the data for the fluted forms (Table 1) indicates that an average of 60 percent of the pebble-sized clasts in these forms are locally derived (up to 4 km up-ice), and the remainder have been transported as far as 10 km. Three bedrock units were selected and utilized as transport indicators on the basis of their position up-ice of many of the linear forms, and the degree to which they could be confidently identified in the pebble fraction. They are: 1) the Piglet Brook rhyolite, 2) a small leucogranite intrusion (Chorlton, 1980), and 3) the Iron Bound intrusives (Dunsworth, 1981) (Figure 2). The most useful of these units is the Iron Bound intrusives, consisting of porphyritic granite and syenodiorite, which are a distinctive bluish-green and greenish-brown, imparted by hornblende, actinolite and

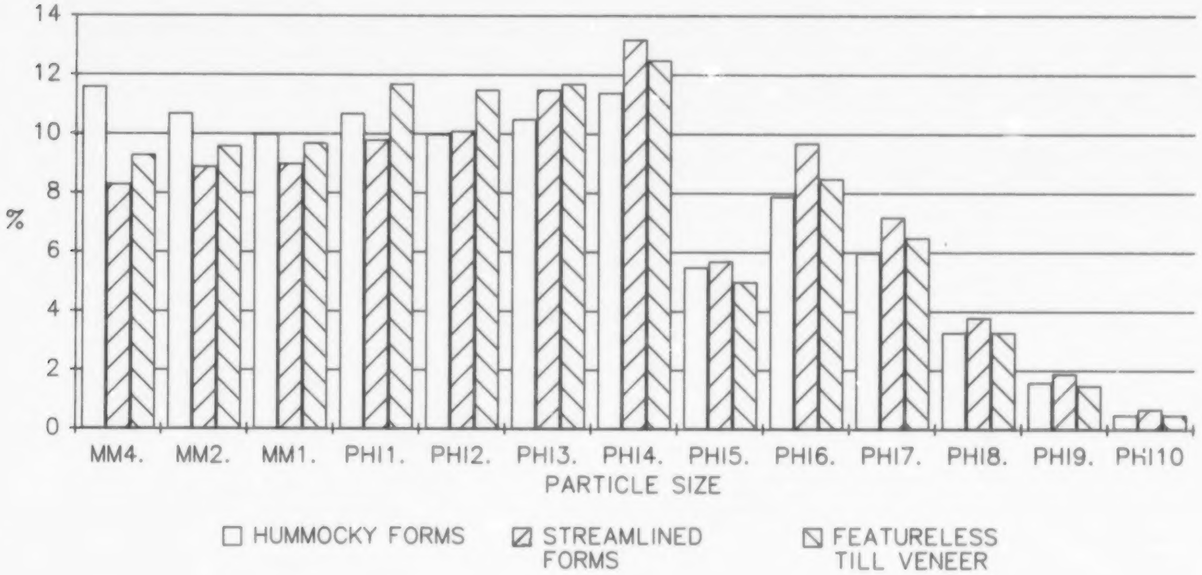

Figure 4. *Mean particle-size distribution by landform type.*

Table 1. Clast provenance in streamlined forms.

Sample	% Derived from up to 4 km up-ice	% More distal sources
1241	54.9	45.1
1242	78.8	21.2
1243	39.9	60.1
1244	48.8	51.2
1245	62	38
1246	64.1	35.9
1103	79.4	20.6
1104	65.6	34.4
1105	83.3	16.7
1107	62.5	37.5
1108	33.3	66.7
1110	50	50
1111	46.5	53.5
1117	45.1	44.9
1118	70	30

biotite. These rocks form a large contiguous area in the central part of the study area, and constitute the highest tills in the area.

Landform Geochemistry and Matrix Texture

The grain-size distributions of the three landforms types are presented in Figure 4. All three are very similar, except that the hummocky forms consist of more coarse to fine sand, and the streamlined forms consist of more fine silts and clay. This difference may simply be a function of the transport distance difference for both these forms.

Pearson correlation coefficients for matrix texture and geochemistry are presented in Table 2. The Pb and Zn geochemical values were regressed with Mn and Fe to remove the effects of co-precipitation, and the Mo values were log transformed (Kerr and Davenport, 1986). The correlation coefficients obtained are uniformly weak, but are statistically significant. A consistent correlation between Mo and the sand content of the matrix is evident in all landform types, and a similar correlation exists between sand and Pb, Zn and Mo in the hummocky forms. A conclusion to be drawn is that some of the variability in Pb, Zn and Mo geochemical values, obtained from regional soil sampling programs, may be explained by these correlations.

Table 2. Pearson correlations; geochemistry and matrix texture by landform type.

	Streamlined forms			Till veneer	Hummocks
	Sand	Silt	Clay	Sand	Sand
Pb				.27	.23
Zn		.22	.22		.22
Mo	.28			.38	.26

Au Distribution in Till (11P/13, 12A/4)

Till samples (-250 mesh) from the Peter Snout (11P/13) and King George IV Lake (12A/4) map areas were analyzed for Au content by fire assay and delayed counting neutron activation. The distribution of Au values greater than 10 ppb, and the occurrence of Victoria Lake Group clasts is shown in Figure 5. Both distributions are spatially similar and thus

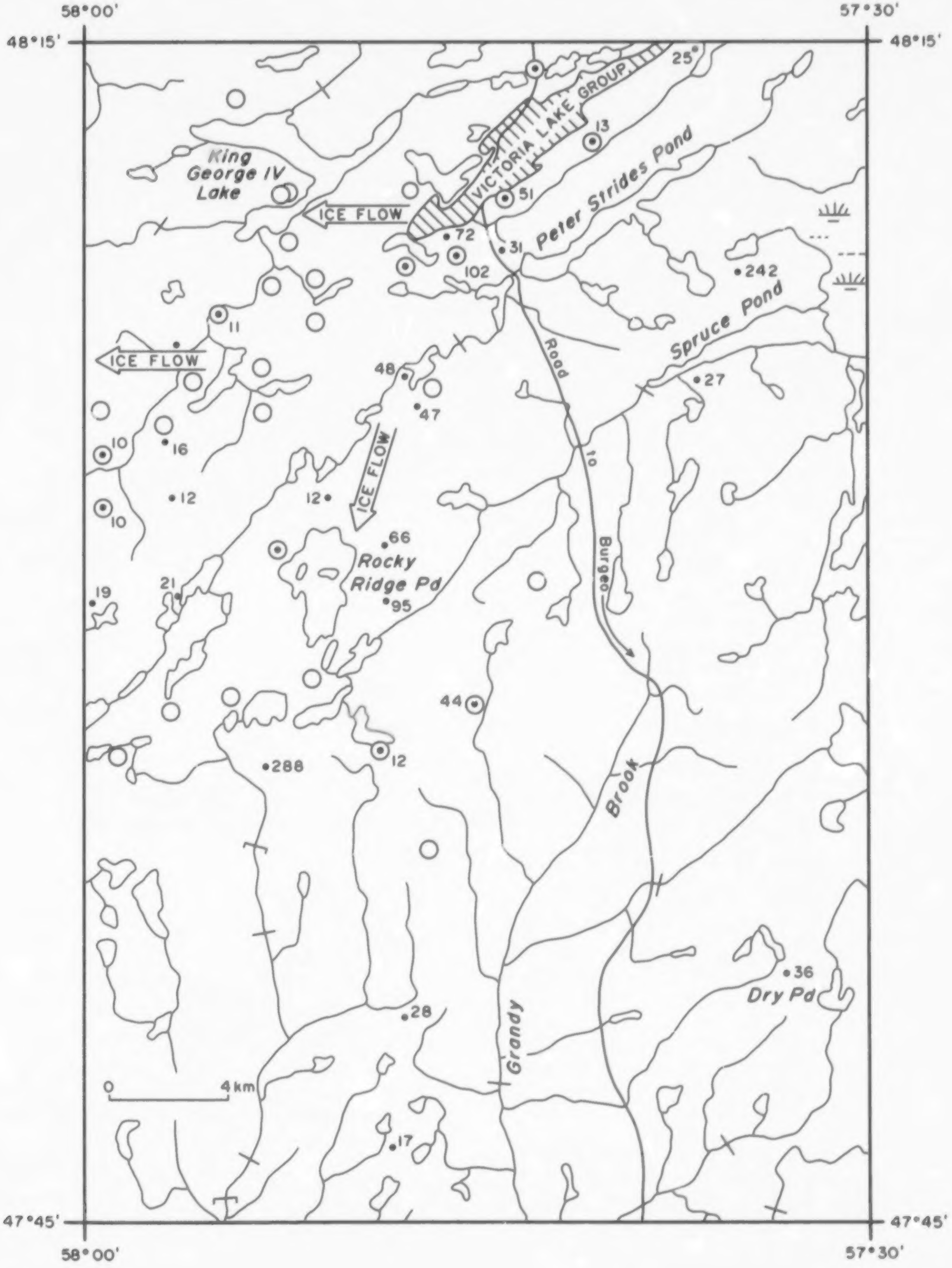
58°00'
57°30'
48°15'
48°15'
25
VICTORIA LAKE GROUP
13
King George IV Lake
Peter Strides Pond
51
ICE FLOW
72
31
102
242
11
Spruce Pond
ICE FLOW
Road
48
27
47
10
16
10
ICE FLOW
12
12
10
66
Rocky Ridge Pd
Burgeo
19
21
95
44
12
288
Brook
36
Dry Pd
28
Grandy
0
4 km
17
47°45'
47°45'
58°00'
57°30'

Figure 5. *Au in till (solid dot) and occurrence of Victoria Lake Group clasts (open circle) 11P/13 and 12A/4. Solid dot in open circle designates both Au in till and occurrence of Victoria Lake Group clasts.*

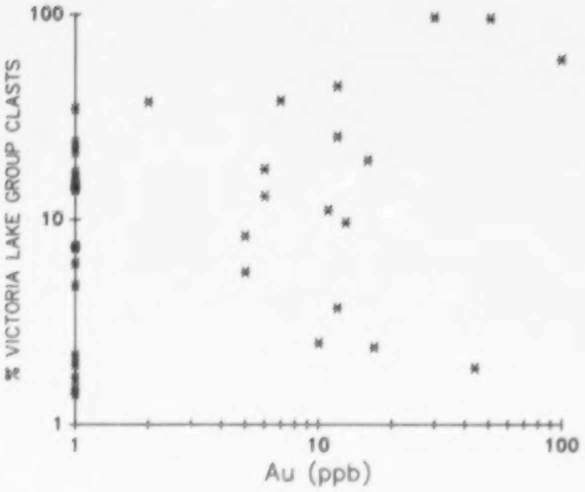

Figure 6. *Scatterplot; Au and percent of Victoria Lake Group clasts.*

indicate that Victoria Lake Group volcanics and sediments may have been the source for the elevated Au values. The Pearson correlation coefficient for Au and clasts in the till (.497) also indicate this source, although the correlation is influenced by several high values (Figure 6). The distribution patterns, which are primarily south and west of the Victoria Lake Group, agree with the known ice flows (south and west) in this area.

CONCLUSIONS

The summarized Quaternary geology and implications for mineral exploration in southwestern Newfoundland are as follows:

1) In much of the area, a southward ice flow dominated, therefore the interpretation of drift anomalies is unlikely to be complicated by complex glacial dispersion;

2) in the northern part of the area, multi-directional flow may have influenced the geochemical signature in the tills;

3) the differing geochemical characteristics of landforms should be considered in the interpretation of soil or till anomalies, where the sampling area includes several landform types;

4) some of the variability in geochemical values obtained in soil sampling programs can be explained by positive correlations between matrix texture and geochemistry;

5) areas of featureless till veneer (no constructional form) provide the best sampling media, and

6) till ramps (pre-crag streamlined forms) contain material from far up-ice and should be sampled only in regional soil surveys.

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MAPPING THE REGIONAL DISTRIBUTION OF GOLD IN NEWFOUNDLAND USING LAKE SEDIMENT GEOCHEMISTRY

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Geochemistry and Geophysics Section

ABSTRACT

The determination of the Au distribution in organic lake sediment is an effective way of establishing the metal's regional distribution in Newfoundland. Although the abundance of Au is low (over 80 percent of samples are below the effective detection limit (2 mg/t) of the direct neutron activation analytical method) and the combined sampling and analytical errors are large, the errors can be diminished by averaging the data to produce smoothed regional distribution maps. The spatial distribution of Au averaged into 7- by 7-km cells, reflects well the main Au prospects of northeastern Newfoundland, and highlights extensive areas of Au enrichment, the largest of which extends over 40 km, north from Gander Lake to Dog Bay, and is now the focus of a very intense exploration program. Several elements are associated with some of the Au showings, e.g., Cr, Cu, Ni, Sb, Sc and Se, reflecting the metal associations of the mineralization and composition of the host rocks. Although no association is universal, Sb has the closest overall relationship, and its regional distribution reflects the surface expression of extensive hydrothermal systems that locally deposited Au mineralization, suggesting that heat from major plutons was the driving mechanism for these systems. The very large size of these hydrothermal systems suggests the possibility of major Au deposits in northeastern Newfoundland.

INTRODUCTION

Mineral exploration in Newfoundland is at record levels, with 1988 expenditures in excess of $40 million, mostly in the search for gold. This exploration effort has been focused, with some success, on two main areas: southwestern Newfoundland, around the Hope Brook Mine and the Cape Ray Prospect, and on the Baie Verte Peninsula, around the former Rambler and Betts Cove mines and prospects in the vicinity of Ming's Bight. These areas represent only two of the many geological environments within the province, some of which may have similar or greater potential for economic gold deposits. To maintain the exploration momentum, the challenge is to find a way to evaluate the gold potential of all favourable geological terrains in the province, thereby providing new exploration targets. As gold deposits worldwide occur in a great variety of geological settings, to be thorough, most of the province should be assessed for its gold potential.

Regional geochemical surveys would be one obvious way to provide this overview for gold, if an effective technique could be developed and demonstrated. Regional lake sediment surveys have been completed for the province at an average sample density of 1 site per 7 km² in Newfoundland and 1 site per 13 km² in Labrador. Thus, to investigate the characteristics of gold dispersion into organic-rich lake sediment, orientation studies around several gold prospects were carried out in 1984 and 1985. These showed that the measurement of gold in organic lake sediment is an effective exploration method, at least at a fairly detailed level, with a sample density of 1 site per 2 km² (McConnell, 1985, 1987; McConnell and Davenport, *in press).*

Further studies have refined the analytical technique used, have evaluated the usefulness of several pathfinder elements, and have demonstrated that the determination of gold in archived, regional lake sediment samples, collected at an average density of 1 site per 7 km², does provide an effective technique for assessing the gold potential of the island of Newfoundland on a regional scale at a very modest cost. In addition, mapping the distributions of Au and related elements, such as Sb, permits the broad scale relationships between these elements and the regional geology to be established, thereby providing unique evidence of the regional metallogeny of Au in Newfoundland. Results are presented, which show the generalized distribution of Au and several related elements in a 30,000-km² area of northeastern Newfoundland (Figure 1). The data have been released at 1:250,000 scale as open files (Davenport and Nolan, 1987, 1988; Davenport *et al.*, 1988).

Sampling and Analytical Methods

Samples of organic-rich, lake-centre sediment (profundal sediment) were collected systematically throughout Newfoundland at an average density of 1 site per 7 km² (Davenport, 1982). The samples were partially air dried in the field, with final drying in the laboratory, in ovens, at 40°C. The dried samples were disaggregated in a mortar and pestle prior to sieving through a 180 μm (80 mesh) stainless-steel sieve, and the fine fraction used for subsequent chemical analysis. Five percent of the samples were randomly selected to be split and included blind as duplicates, to monitor analytical precision.

This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

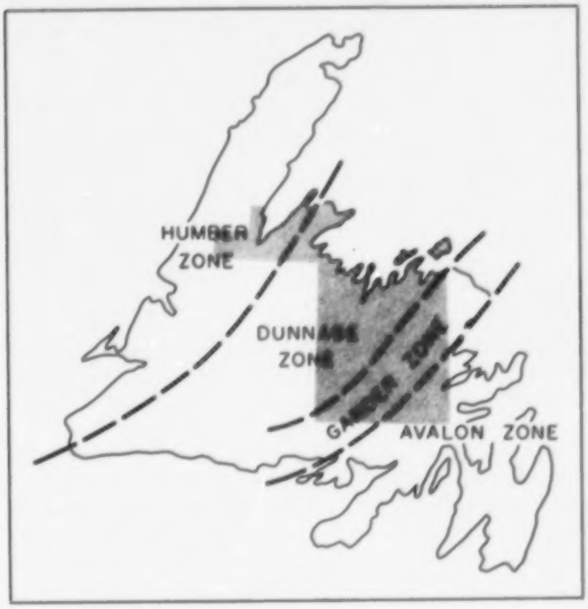

Figure 1. *Index map showing the area of northeastern Newfoundland in which the regional distribution of gold has been mapped using lake sediment geochemistry.*

Direct instrumental neutron activation analysis was used to determine Au, Sb, As, Se, W and Cr (together with Ba, Br, Ce, Cs, Eu, Hf, La, Lu, Rb, Sm, Sc, Na, Ta, Tb, Th, U, Yb and Zr, none of which are discussed in this paper). The average sample weight used was 10 g, but the actual weight varied from sample to sample, depending on the amount and density of material available. The characteristics of the results obtained by this method were discussed by Davenport (1988).

Atomic absorption spectrophotometry was used to determine Cu, Pb, Zn, Co, Ni, Ag, Fe and Mn following the digestion of 1 g of sample in a 4M HNO_3- 0.1 M HCl acid mixture. Loss-on-ignition (LOI) was determined as a measure of organic carbon content by ashing the samples during 3-hour-controlled temperature rise to 500°C. These techniques were described by Wagenbauer *et al.* (1983).

REGIONAL GEOCHEMICAL DISTRIBUTIONS

Enhancement of Regional Geochemical Variation

Gold in almost all geological materials, and at most scales, is characterized by low abundances, but a high proportion of the observed variation is due to sampling and analytical noise. The sampling noise is inherent for most media because of the nugget effect, i.e., the occurrence of gold as a major constituent of a very minor phase. Analytical precision can be improved by adopting more precise (and costly) analytical techniques, and sampling errors can be diminished by careful sample preparation and the use of large sample weights for analysis (also costly) but there are economic limits as to how far these steps can be taken.

The results discussed here are from archived samples where only a limited amount of sample material is available (generally about 10 g), and sample preparation had already been carried out. Fortunately, in organic-rich lake sediment gold has a relatively minor sampling variance when compared with many other media, but nonetheless its distribution in these lake sediments is much noisier than most of the other trace elements determined. This is illustrated in Figure 2 where replicate values for Au and Sb in site duplicates from northeastern Newfoundland are shown. Whereas Sb reproduces well (r = 0.86), Au does not (r = 0.34), at least not below about 5 mg/t. An analysis of variance to compare between site to within site variance (Garrett, 1973) reveals as expected that most of the variance of Sb occurs between sites, whereas for Au most of the variance is between site duplicates. Part of the problem with Au is that over 80 percent of the values are below the nominal analytical detection limit of 2 mg/t, and most of the rest of the data is only slightly higher (2 to 4 mg/t), where analytical precision is poor (± 100 percent).

In areas of Au enrichment, the site duplicate data for Au are much more reliable (McConnell, 1987), and the distribution of Au in lake sediment around known Au prospects suggests that anomalous Au values in lake sediment are both reproducible and geologically meaningful (McConnell, 1985, 1987; McConnell and Davenport, *in press).* On a regional scale, the Au data in lake sediment are too noisy to contour. When presented as symbol plot maps (Figure 3), however, fairly coherent patterns of anomalous Au values can be distinguished from areas of consistently low values, suggesting that there are real patterns in the data. Averaging Au values from several adjacent sites should improve the signal to noise ratio if indeed there are meaningful patterns present that are obscured by random sampling and analytical noise.

To emphasize regional geochemical patterns in Newfoundland, the geochemical data from individual lake sites have been averaged within 7- by 7-km cells (Davenport, 1982). This cell size was found to yield smoothed trace-element patterns that could be correlated well with the regional geology and the distribution of mineral deposits, and has therefore been adopted here. Because element distributions are positively skewed, geometric cell-means were calculated to diminish the effect of high outliers. The median could not be used for severely truncated data such as Au, as most cell medians were below the detection limit of 2 mg/t. The absolute value of the cell-means for Au is dependent on the value assigned to samples with a Au content below the detection limit (in this study, an abitrary value of 1 mg/t was used). The values for the cell-means for Au are not, therefore, accurate averages of actual Au abundance in lake sediment within each cell (they are probably too high), but they do reflect relative changes in average abundances from cell to cell.

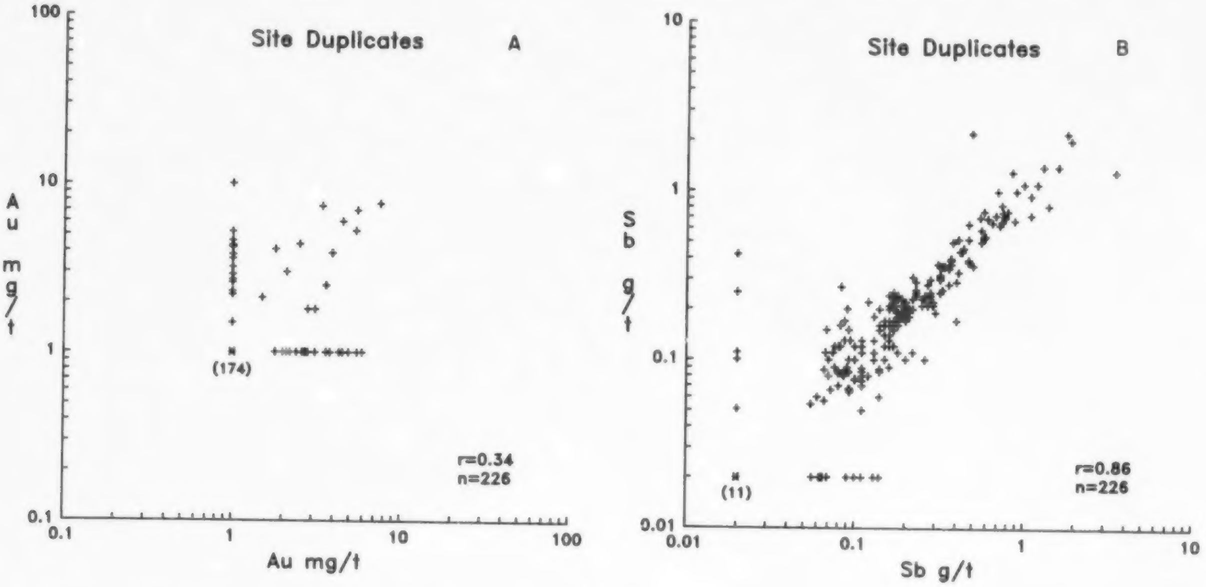

Figure 2. *Scatterplots of Au (A) and Sb (B) in site duplicate samples of lake sediment from northeastern Newfoundland.*

One indication that averaging the geochemical data has increased the geochemical signal relative to noise in the Au data is provided by the correlations between it and the other elements. Correlation matrices for the individual site data and the cell-averaged data is presented in Table 1. Correlations between Au and Sb, Cu, Ni, Cr, Se and Sc are all higher in the averaged data than in the site data, although there is little or no overall increase in the values of the correlation coefficients between elements in the cell-averaged data compared with the site data (Figure 4).

Distributions of Gold and Antimony

The spatial distributions of Au and Sb in the cell-averaged data are shown in Figures 5 and 6 in relation to the regional geology (after Williams *et al.*, 1988), and the main Au prospects in northeastern Newfoundland (from Tuach *et al.*, 1988). Contour intervals were chosen from cumulative frequency plots. For Au, there is a break in slope at the 87th percentile (1.6 mg/t), with about 5 percent of the cells having >2.0 mg/t. All of the significant Au prospects, except the Kim Lake showings (Occurrence 16 in Figures 5 and 6), coincide with Au in lake sediment values above the 69th percentile (>1.35 mg/t), and in most cases above 1.6 mg/t. The lack of response in Au in the Kim Lake area (Occurrence 13, Figures 5 and 6) is not understood–the prospects are, however, reflected by anomalous Sb levels. The most interesting feature of the regional Au distribution is the extensive area of high Au values between Gander Lake and the coast, in the northeast part of the area. This area is currently where the most active preliminary Au exploration in the province is taking place.

Antimony shows a change in background levels, increasing from west to east. Although not obvious, Sb does reflect well the known Au prospects (Occurrences 1 and 2, Figures 5 and 6) on the west side of White Bay and on the Baie Verte and Springdale peninsulas (Davenport and McConnell, 1988), but only relative to the local background. The highest Sb values are found in the eastern part of the area, in the south around the Kim Lake Au–Sb showings, northeast of Gander Lake, around the margins of the Mount Peyton gabbro–granodiorite pluton, and in a belt extending northeast from this intrusive body to Gander Bay. Most of these areas of high Sb in lake sediment are areas of active Au exploration programs, some initiated prior to the release of these data, but several subsequent to their release.

DISCUSSION AND CONCLUSION

The regional Au distribution in organic lake sediment does reflect the known Au prospects in northeastern Newfoundland, and perhaps more importantly, indicates extensive areas of Au enrichment that are only beginning to receive exploration interest. Several elements (As, Ba, Cr, Cu, Ni, Pb, Sb, Sc and Zn) are locally associated with Au in lake sediment, reflecting either the mineralization itself, or the composition of the host rocks to mineralization, with Sb being the single most generally useful pathfinder element in Newfoundland. Antimony reflects all the more significant gold prospects in northeastern Newfoundland, including the Kim Lake showing, which is not reflected by Au. In the Port aux Basques area, Sb was similarly effective in reflecting the Hope Brook and Cape Ray deposits, where again Au was not (Davenport and McConnell, 1988).

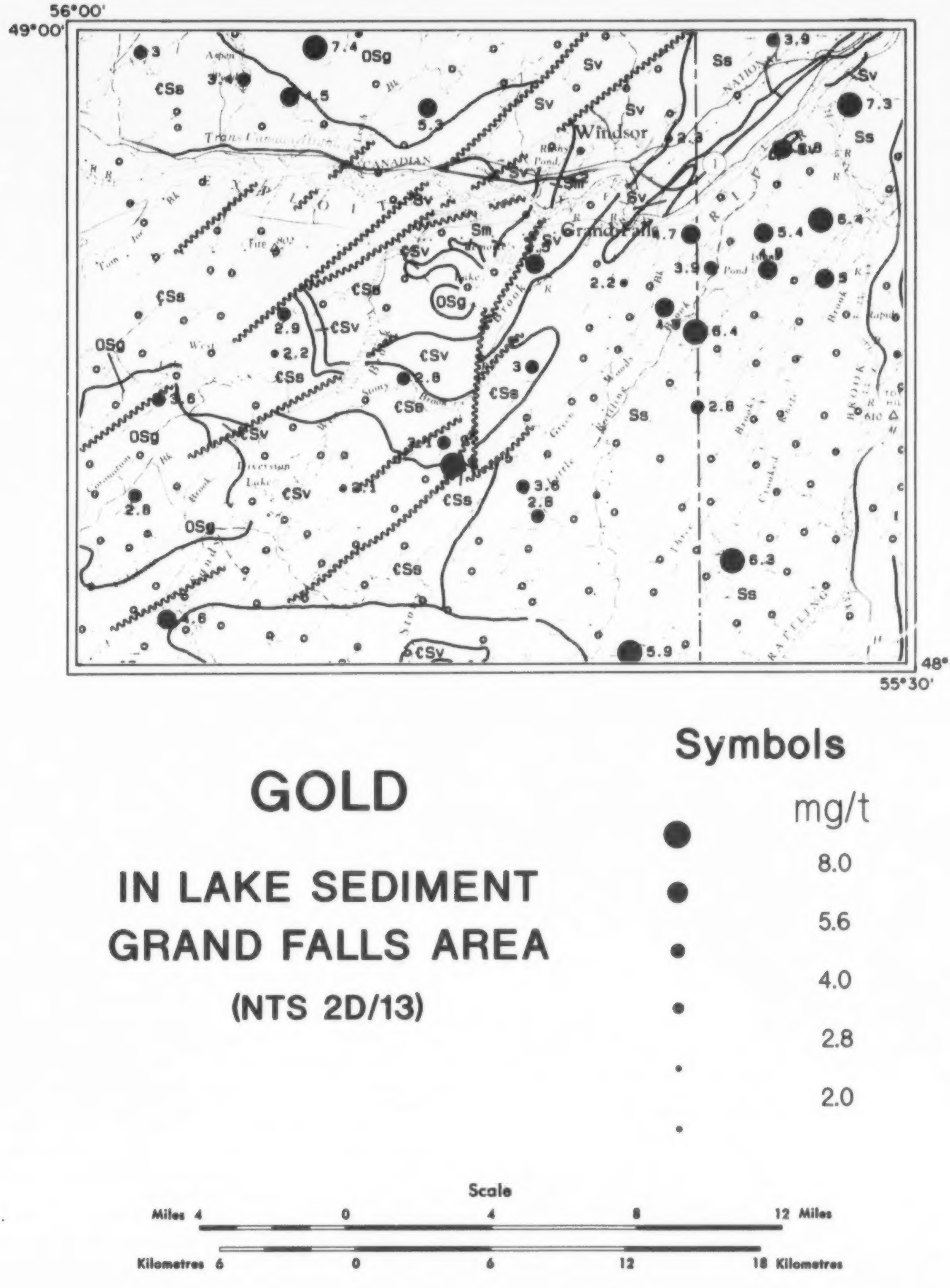
56°00'
49°00'
48°45'
55°30'
Windsor
Grand Falls
OSg
CSs
CSv
Sv
Ss
Sm
7.4
4.5
5.3
3.9
7.3
2.3
5.4
3.9
2.2
2.9
2.2
6.4
2.8
3.8
2.8
2.8
6.3
5.9
GOLD
IN LAKE SEDIMENT
GRAND FALLS AREA
(NTS 2D/13)
Symbols
mg/t
8.0
5.6
4.0
2.8
2.0
Scale
Miles 4
0
4
8
12 Miles
Kilometres 6
0
6
12
18 Kilometres

Figure 3. *Gold in lake sediment in the Grand Falls area, Newfoundland, presented as a symbol plot map.*

Table 1. Matrix of correlation coefficients for Au, Sb, As, Cu, Ni, Cr, Se, Ag, W, Ba, Sc, Pb, Mn, Fe, LOI, lake area and lake depth in individual sample sites (upper right, n=4023) and cell averaged data (lower right, n=685) in northeastern Newfoundland. Coefficients not significant at the 99 percent confidence level are shown as a dot.

SITES

	Au	Sb	As	Cu	Ni	Cr	Se	Ag	W	Ba	Sc	Pb	Mn	Fe	LOI	
Au		8	.	12	11	9	12	.	.	.	9	.	.	.	.	Au
Sb	13		52	21	40	39	7	-10	.	32	41	11	32	18	-17	Sb
As	.	57		.	37	31	9	-9	23	24	21	11	56	54	-27	As
Cu	19	18	.		59	39	22	27	-14	21	60	7	10	.	32	Cu
Ni	22	42	37	62		64	.	11	.	33	49	16	27	16	.	Ni
Cr	14	38	30	45	75		.	.	10	44	66	16	33	30	-12	Cr
Se	16	14	.	23	.	.		.	.	.	16	-8	10	9	16	Se
Ag	.	-16	.	15	.	.	.		.	11	9	16	.	.	14	Ag
W	.	.	20	-29	.	.	.	.		10	.	6	18	24	-22	W
Ba	12	23	.	27	31	38	.	13	.		56	22	44	42	-19	Ba
Sc	15	34	.	70	49	59	21	.	-18	50		6	39	35	.	Sc
Pb	.	.	.	.	19	19	.	27	.	27	.		25	20	-10	Pb
Mn	.	34	53	.	21	25	.	.	20	35	27	32		81	-27	Mn
Fe	.	.	39	.	.	18	.	.	30	35	22	27	76		-35	Fe
LOI	.	-18	-31	41	.	.	22	.	-31	.	.	.	-33	-40		LOI
	Au	Sb	As	Cu	Ni	Cr	Se	Ag	W	Ba	Sc	Pb	Mn	Fe	LOI	

CELLS

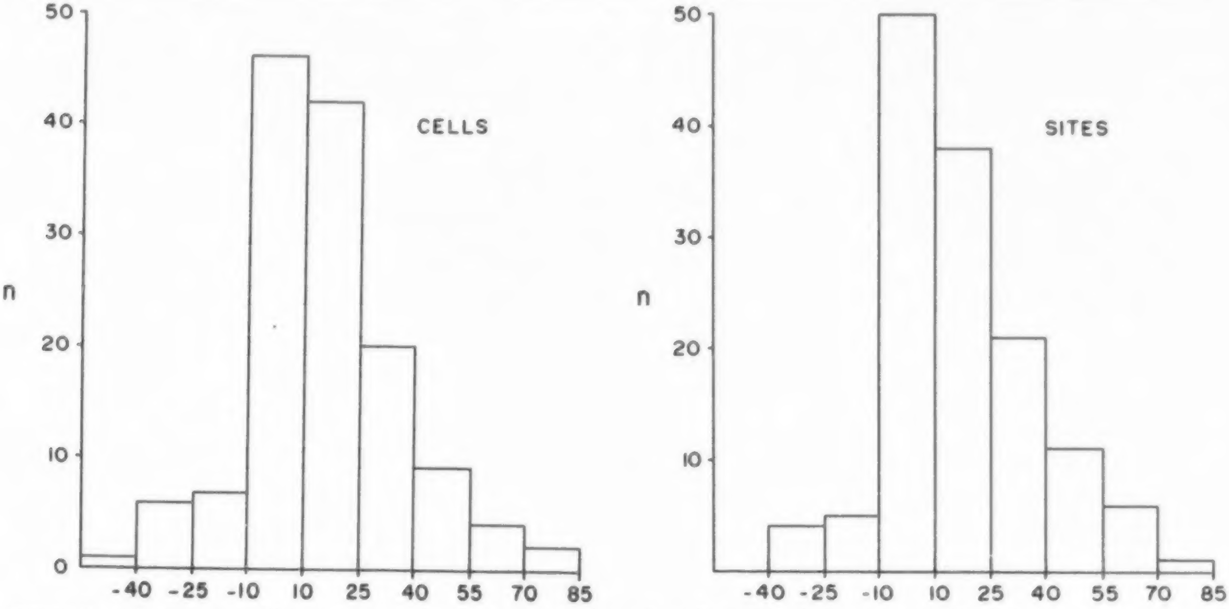

Figure 4. *Histograms of inter-element correlation coefficients in cell-averaged data and site data for lake sediments in northeastern Newfoundland.*

Antimony is distributed in an ovoid pattern surrounding the Mount Peyton gabbro–granodiorite pluton, a pattern that is extended along major, regional faults that trend northeast-southwest, away from the pluton. The pattern of Sb enrichment is developed over both marine (oceanic) sedimentary sequences as well as predominantly subaerial clastic rocks, and its form suggests that the pluton may have been the heat source that drove the hydrothermal system,

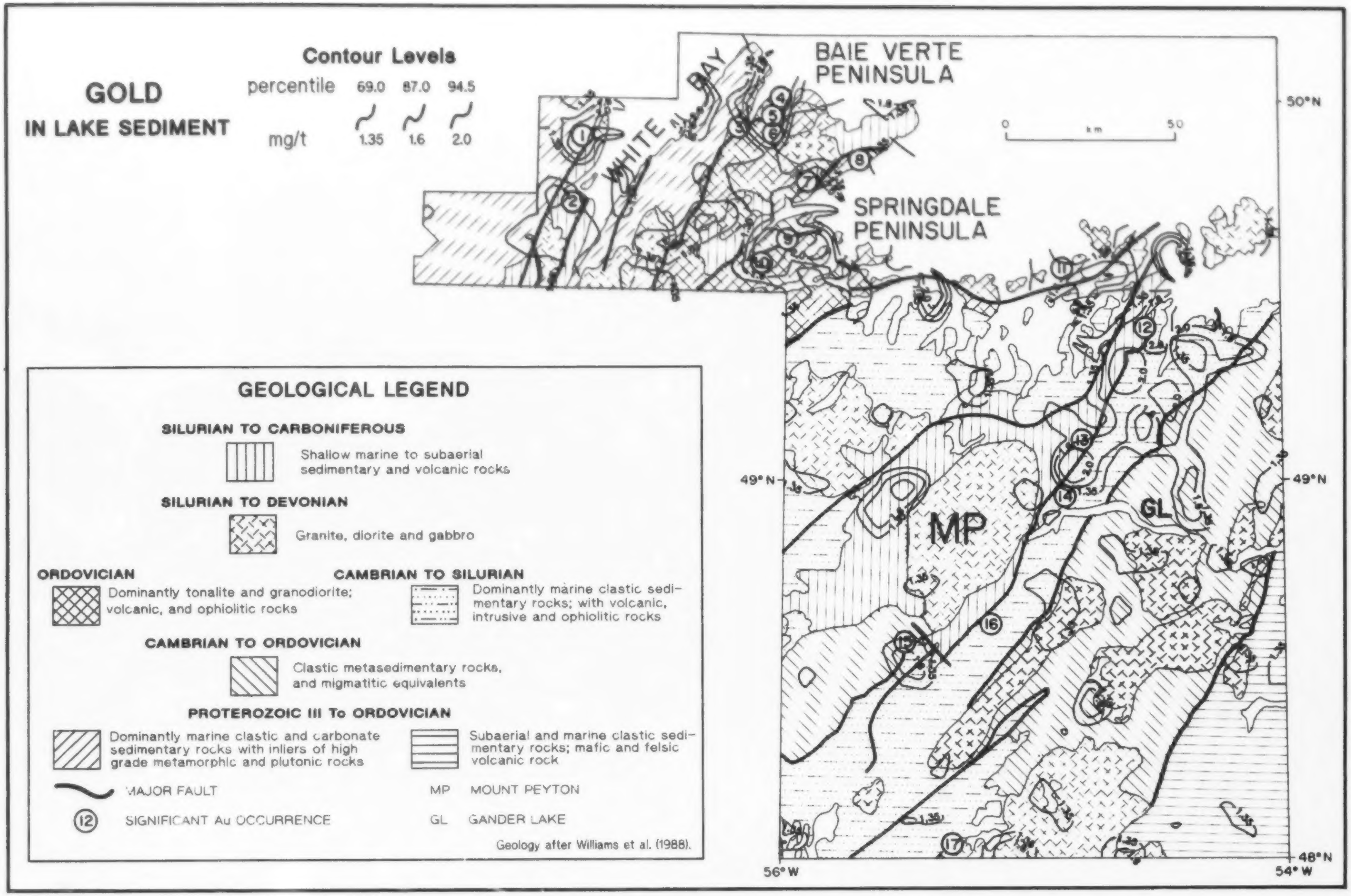
GOLD
IN LAKE SEDIMENT
Contour Levels
percentile 69.0 87.0 94.5
mg/t 1.35 1.6 2.0
BAIE VERTE PENINSULA
SPRINGDALE PENINSULA
WHITE BAY
MP
GL
0 km 50
50°N
49°N
48°N
56°W
54°W
GEOLOGICAL LEGEND
SILURIAN TO CARBONIFEROUS
Shallow marine to subaerial sedimentary and volcanic rocks
SILURIAN TO DEVONIAN
Granite, diorite and gabbro
ORDOVICIAN
Dominantly tonalite and granodiorite; volcanic, and ophiolitic rocks
CAMBRIAN TO SILURIAN
Dominantly marine clastic sedimentary rocks; with volcanic, intrusive and ophiolitic rocks
CAMBRIAN TO ORDOVICIAN
Clastic metasedimentary rocks, and migmatitic equivalents
PROTEROZOIC III To ORDOVICIAN
Dominantly marine clastic and carbonate sedimentary rocks with inliers of high grade metamorphic and plutonic rocks
Subaerial and marine clastic sedimentary rocks; mafic and felsic volcanic rock
MAJOR FAULT
⑫ SIGNIFICANT Au OCCURRENCE
MP MOUNT PEYTON
GL GANDER LAKE
Geology after Williams et al. (1988).

Figure 5. *Distribution of Au in organic-rich lake sediment in northeastern Newfoundland in relation to the regional geology and main Au prospects: 1) Apsy Zone; 2) Unknown Brook prospect, Browning Mine; 3) Terra Nova Mine; 4) Deer Cove Prospect; 5) Stoger Tight prospect, Goldenville Mine; 6) Rambler Mines; 7) Betts Cove Mine, Nugget Pond Prospect; 8) Tilt Cove Mine; 9) Little Bay Mine, Mine Brook prospect; 10) Randell–Jackman prospect; 11) Moretons Harbour prospects; 12) Clutha showing; 13) Blue Peter showing; 14) Mustang, Bullet and Bowater showings; 15) Great Bend prospect; 16) Aztec and Goose showings; 17) Kim Lake prospects.*

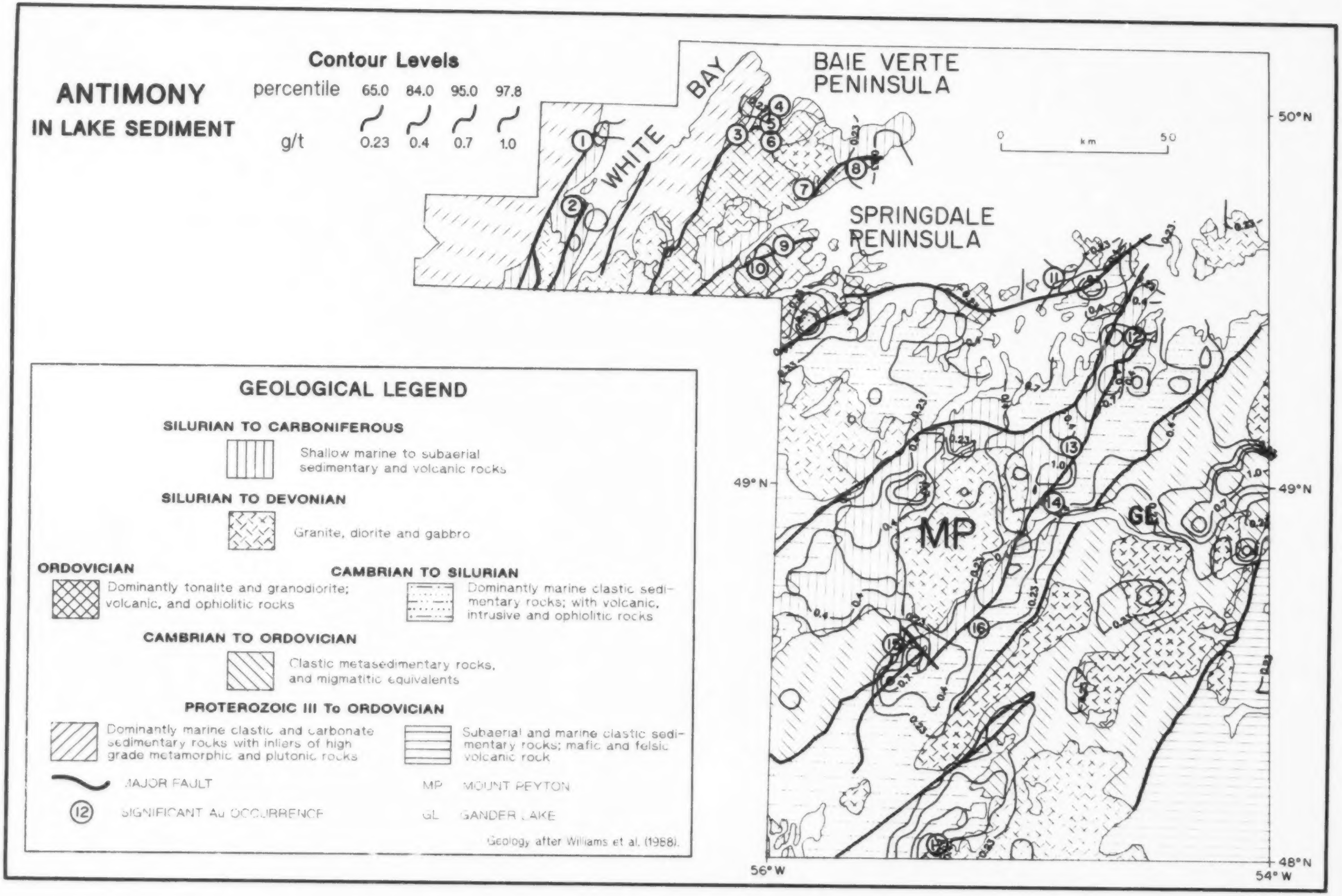
ANTIMONY
IN LAKE SEDIMENT
Contour Levels
percentile 65.0 84.0 95.0 97.8
g/t 0.23 0.4 0.7 1.0
0 km 50
WHITE BAY
BAIE VERTE PENINSULA
SPRINGDALE PENINSULA
MP
GL
50°N
49°N
48°N
56°W
54°W
GEOLOGICAL LEGEND
SILURIAN TO CARBONIFEROUS
Shallow marine to subaerial sedimentary and volcanic rocks
SILURIAN TO DEVONIAN
Granite, diorite and gabbro
ORDOVICIAN
Dominantly tonalite and granodiorite; volcanic, and ophiolitic rocks
CAMBRIAN TO SILURIAN
Dominantly marine clastic sedimentary rocks; with volcanic, intrusive and ophiolitic rocks
CAMBRIAN TO ORDOVICIAN
Clastic metasedimentary rocks, and migmatitic equivalents
PROTEROZOIC III To ORDOVICIAN
Dominantly marine clastic and carbonate sedimentary rocks with inliers of high grade metamorphic and plutonic rocks
Subaerial and marine clastic sedimentary rocks; mafic and felsic volcanic rock
MAJOR FAULT
MP MOUNT PEYTON
⑫ SIGNIFICANT Au OCCURRENCE
GL GANDER LAKE
Geology after Williams et al. (1988).

Figure 6. *Distribution of Sb in organic-rich lake sediment in northeastern Newfoundland in relation to the regional geology and main Au prospects (see Figure 5 for prospect names).*

which localized the Au mineralization. If this is so, it would represent a very major system involving very large fluid volumes affecting huge rock volumes, and which may well have produced large Au deposits.

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LAKE SEDIMENT AND WATER GEOCHEMICAL SURVEYS FOR RARE-METAL MINERALIZATION IN LABRADOR

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ABSTRACT

A study was undertaken, a) to determine the feasibility of applying new analytical procedures for determining elements of the rare-metal suite to conventional lake sediment—water surveying and, b) to identify new prospecting targets. Four areas were selected for surveying using available reconnaissance lake sediment data for elements commonly associated with rare-metal mineralization, such as fluorine, zinc and lead. The presence of rare-metal mineralization was also a factor for selecting the Strange Lake and Letitia Lake areas. A total of 444 pairs of sediment and water samples were collected from a 2080-km^2 area and analyzed for up to 38 elements. All sediment samples were analyzed by Inductively Coupled Plasma Emission Spectrometry (ICP) for: Ba, Ce, Ga, La, Nb, Sr, Th, Y, and Zr, and by Atomic Absorption Spectrometry (AA) for Ag, Be, Cd, Co, Cu, Fe, Mn, Mo, Ni, Pb and Zn. Fluoride was determined by ion-selective electrode and loss-on-ignition as an estimate of organic content. Samples from the Strange Lake area were also analyzed by direct Neutron Activation Analysis (NA) for: As, Au, Ba, Br, Ce, Co, Cr, Cs, Eu, Fe, Hf, La, Lu, Mo, Na, Ni, Rb, Sb, Sc, Sm, Ta, Th, U, W, Yb, Zn, and Zr. All lake waters were analyzed for F, Zn and pH. Those from the Letitia lake area were also analyzed for Y and water samples from the Strange Lake area were analyzed for Be.

The use of correlation analysis, principal components analysis and simple statistics helped in recognizing associations of elements in each area. Cumulative frequency curves are employed to recognize and separate component populations and to define contour intervals for contoured data maps of Be, Yb and Y in sediment, and F in water. The rare-metal suite and associated elements (e.g., REE, Be, F, Nb, Pb, Th, Y, Zn and Zr) give clearly defined dispersion patterns upward of 40 km in length at Strange Lake, which can be linked to areas of known mineralization. In addition, several anomalies in areas that have received little or no exploration, present new targets for exploration. Contoured data for F, Zn and Y in water, show dispersion patterns that can be linked with mineralization and/or areas of alkaline bedrock.

INTRODUCTION

This paper summarizes data and observations contained in a recently published open file report (McConnell, 1988) on the results of field work during 1983 and 1985, following the discovery of the Strange Lake (Lac Brisson) rare-metal deposit in Labrador. This type of deposit is receiving exploration attention from industry, with the expectation that demand for these commodities could rapidly increase in the near future. The study characterizes the nature of geochemical dispersion into the surficial environment; the emphasis being on lake sediment and water geochemistry. It also assesses the mineral potential of four areas with known or suspected rare-metal mineralization, and identifies several elements, not previously analyzed in lake sediment, as useful guides to mineralization. These include the rare-earth elements, Be, Nb, Y and Zr. Analyses of Y in water are also easily performed by ICP and correlate with yttrium mineralization.

The Strange Lake and Letitia Lake areas, where rare-metal mineralization is present, were sampled. A 50-million-tonne deposit of Zr−Y−Be−Nb−REE is contained within an alkaline intrusion at Strange Lake (Miller, 1988) and several smaller Nb−Be and Zr occurrences are located in the Letitia Lake area, generally in association with peralkaline rocks (Miller, 1988). The Letitia Lake area has anomalous levels of F and Zn (Geological Survey of Canada, 1983). The two other areas studied, referred to here as Michikamats and 13L, were not known to be mineralized but were surveyed because the presence of very high levels of F in the reconnaissance data (Geological Survey of Canada, 1978a,b, 1983) suggested the occurrence of specialized or possibly mineralized granitoids. The locations of these four areas is shown in Figure 1.

PREVIOUS GEOCHEMICAL WORK

The earliest work is that of Brummer (1957, 1960), who described stream sediment surveys for base metals in the Letitia Lake area that led to the discovery of the Mann Nb−Be deposits. In 1977, the Geological Survey of Canada released the first of a series of open file reports of reconnaissance lake sediment and water survey data for Labrador. The samples were not analyzed for any of the rare-metal suite ore elements, but did include elements commonly associated with such deposits including F, Pb, U and Zn. The surveys involved the collection of samples at a density of about 1 per 17 km^2. Subsequent open file reports (Geological Survey of Canada, 1978a,b) included the Strange Lake area, the Michikamats area and the 13L areas, and highlighted several areas with

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This project is a contribution to the Canada−Newfoundland Mineral Development Agreement, 1984−1989

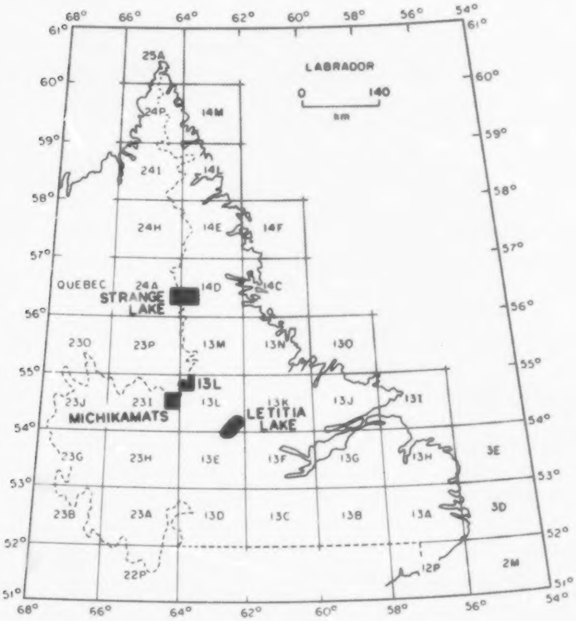

Figure 1. *Location of survey areas.*

anomalous concentrations of F, Pb, U and Zn. One of the anomalies was instrumental in leading to the discovery of the Strange Lake deposit. In 1983, reconnaissance coverage was extended to the Letitia Lake area, in which anomalies in Pb, F and Zn were identified. Work by McConnell and Batterson (1987) on the geochemistry of tills, stream and lake sediments and waters in the Strange Lake area indicated the presence of strong linear anomalies in all sampled environments, extending for tens of kilometres down-ice from the deposit. The lake data in the report did not include analyses of such diagnostic elements as rare earths, Nb, Th, U, Y or Zr, a gap which is filled in this report.

AREA DESCRIPTIONS

Strange Lake Area

The area is located on the Labrador–Quebec border and is accessible only by air. The regional geology of the area has been mapped by Ryan *et al.* (1988) and the deposit occurs in the Helikian Strange Lake alkaline complex, the lithology and mineralogy of which has been comprehensively described by Miller (1986). The alkaline complex, which intrudes Aphebian gneiss and a Helikian quartz monzonite body, has an ovoid surface exposure. The most intensely mineralized phase has a surface exposure of about 1.5 km². The ore mineralogy in the 50-million-tonne deposit is complex and consists primarily of such unusual minerals as elpidite (Na–Zr silicate), gittinsite (Ca–Zr silicate), pyrochlore (Ca niobate), and gadolinite (Y–Be silicate).

The surveyed area is located above the tree-line and has a low to moderate topographic relief. Eastward-flowing glacial ice of the Laurentide ice-sheet, which covered the area in the Late Wisconsinan period (Shilts *et al.*, 1979; Batterson *et al.*, 1985), eroded the upper portion of the alkaline complex and deposited till over 60 percent of the area, ranging in thickness from a veneer to 16 m. The remainder of the surface is covered with glaciofluvial deposits, organic accumulations and scattered outcrops (McConnell and Batterson, 1987).

Letitia Lake Area

This area is readily accessible by air from Goose Bay, Labrador. Bedrock mapping by Thomas and Hibbs (1983) and Thomas *et al.* (1983) resulted in a five-fold division of the rocks in the survey area; the oldest of which is the Aphebian Disappointment Lake gneiss. This is intruded by the dioritic to granitic North Pole Brook intrusive suite. Overlying these are felsic volcanic and subvolcanic rocks and associated sedimentary rocks of the Letitia Lake Group, which have peralkaline affinities (Thomas, 1981). These, in turn, have been intruded by the Red Wine alkaline intrusive suite. Finally, the four older rock units are unconformably overlain by sedimentary, mafic volcanic and intrusive rocks of the Seal Lake Group.

Most of the known Be–Nb–REE mineralization in the vicinity of Letitia Lake is associated with peralkaline syenite of the Red Wine alkaline intrusive suite (Miller, 1987). Miller reports that the principal Nb–Be-bearing minerals are barylite (Be), eudidymite (Be), niobophyllite (Nb) and pyrochlore (Nb). The most notable occurrences are the Mann #1 and Mann #2 deposits (Be–Nb–Th), and the Two Tom Lake occurrence (Be–Y–Nb–REE). The Mann #1 deposit is estimated to contain 6.8-million tonnes of BeO (Gaines, 1977). Widespread, but apparently minor, Zr mineralization is found south of Letitia Lake in eudialyte-bearing pegmatites, also thought to be associated with the Red Wine alkaline intrusive suite (R. Miller, personal communication, 1988).

Work by Batterson and LeGrow (1986) indicated that at least three major glacial flow events occurred during the Wisconsinan. The earliest trends 046°; the second ranges from 096° in the south, to 114° in the north. The third and last event, however, is the major flow in terms of controlling the present glacial dispersion patterns of both geochemistry and lithic clasts. Its flow direction ranges from 074° in the north, to 080° in the south. Most of the area is underlain by locally derived till, which is generally < 4-m thick. Other locally significant Quaternary deposits mapped by Batterson and LeGrow (1986) include glaciofluvial outwash and bog.

Michikamats Area

This area is accessible by float plane or helicopter from Goose Bay or Schefferville. The geology in the area has been mapped at 1:250,000 scale by Emslie (1963), and is summarized here.

The oldest rocks, which are located in the northwest, are fine- to medium-grained, foliated granite and granodiorite. Massive, medium- to coarse-grained, pink to white granite and syenite underlie most of the central part of the area. A subcircular, 10-km diameter of clinopyroxene-bearing syenite represents a distinct phase of this unit. The youngest rocks consist of flat-lying, unmetamorphosed conglomerate, arkose and siltstone and are poorly exposed in the southeast. No mineralization of economic interest was noted by Emslie (1963) during the regional mapping.

Detailed Quaternary mapping has not been done in the Michikamats area. Emslie (1963) noted during the course of his bedrock mapping that drumlins, striae, and crag and tail features indicate that the last glacial advance was east to east-southeast. He also noted rare striae, which he thought recorded an earlier northeast advance. Boulder till covers most of the area and outcrops are restricted mainly to hilltops and along lake shores. Several eskers are present in the area trending southeast and cross-cutting the dominant striae direction. One of these eskers traverses the syenite phase described above and passes over the sedimentary terrane. It appears to play a significant role in the dispersal and localization in lake sediments and waters of some elements derived from the syenite.

13L Area

The 13L area formed a small part of the region, mapped at 1:250,000 by Emslie (1964). The oldest rocks underlie the central survey area and consist of granitic and/or syenitic gneiss metamorphosed to granulite facies. The granitic gneiss terrane in the south is a porphyritic granitic gneiss and, in fact, may be a phase of the previous unit (Emslie, 1964). Greene (1970) regarded the first unit as Aphebian and the porphyritic rock as Paleohelikian. The youngest rocks are massive, medium- to coarse-grained, granitic rocks (Emslie, 1964). Scintillation counts over this last unit average 1.5 to 3 times those found in other granitoid terranes (McConnell, 1986). No mineralization has been reported from the area.

The only Quaternary geology observations reported are those of Emslie (1964), and are included in the marginal notes of his geology map. He describes abundant glacial striae throughout the map area, which indicate the latest ice-flow direction to have been easterly, however, within the present survey area only a single ice-striation having a 100° trend is recorded.

SAMPLING AND ANALYTICAL METHODS

Lake sediment and water samples were collected by means of a float-equipped helicopter; 444 samples, including 5 percent site duplicates were obtained from 407 lakes in the four survey areas. Sample density was 1 site per 4.7 km±The sampled lakes have a mean (geometric) area of 0.11 km±and a mean ± 2 standard deviation range of 0.01 km±to 2.88 km± Details of sample collection and sample preparation are described in McConnell (1988). Sediments were dried, sieved through a 180 μm mesh size and analyzed for up to 38 elements.

Inductively coupled plasma emission spectroscopy (ICP) was used to analyze all sediment samples for Ba, Ce, Ga, La, Nb, Sr, Th, Y, and Zr. The analyte was prepared by digesting one gram of sample overnight in 15 mL concentrated HF, 5 mL concentrated HCl and 5 mL of 50 percent HClO#. This solution was evaporated to dryness at 200° C. Ten mL of concentrated HCl and 40 mL of water were then added. This mixture was redigested at 100° C for 2 hours, brought to volume by adding 20 percent HCl and centrifuged to remove any remaining solids. The solution was then analyzed by ICP. This digestion gives complete solution of all elements determined except, in some cases, Zr.

Samples were prepared for atomic absorption spectrometry (AA) as follows: 1 g was added to 6 mL of 4 M HNO^{a} and 1 M HCl, allowed to remain overnight, then mixed and digested for two hours at 90° C, cooled, and finally, made up to 20 mL by adding de-ionized water. This solution was then analyzed for Ag, Be, Cd, Co, Cu, Fe, Mn, Ni, Pb and Zn. This digestion typically extracts 80 to 100 percent of these elements (Davenport, 1988). The preparation and analysis of Mo by AA is described in Wagenbauer *et al.* (1983).

Fluoride was determined by ion-selective electrode, and loss-on-ignition was measured as an estimate of organic content. Samples from the Strange Lake area were also analysed by direct Neutron Activation Analysis (NA) for: As, Au, Ba, Br, Ce, Co, Cr, Cs, Eu, Fe, Hf, La, Lu, Mo, Na, Ni, Rb, Sb, Sc, Sm, Ta, Th, U, W, Yb, Zn, and Zr. All the lake water samples were analyzed for F, Zn and pH. Those from the Letitia Lake area were also analyzed for Y by ICP. Waters from the Strange Lake area were analyzed for Be by AA, although nearly all samples were below the 10 μg/L detection limit.

RESULTS

A variety of interpretation procedures were used to obtain statistics for inter-area comparisons and to determine correlations between elements. These aided in the identification of environmental factors that affect the compositions of sediment, enabled overlapping geochemical populations to be separated into their components, and facilitated the recognition of distinct groups of elements associated with rare-metal mineralization. Cumulative frequency plots are of particular help in defining threshold levels for sediment concentrations derived from mineral occurrences. Other procedures include the computation of basic statistics (means, standard deviations and ranges), histograms, Pearson correlations, regression analysis and principal components analysis.

Principal components analyses is a multivariate statistical technique that defines groups of variables (elements) that are mutually correlated, and, for each group, defines a 'factor' that is a linear combination of its variables that accounts for most of its variance. In geochemistry, it is particularly useful in identifying groups of associated elements that are separated into factors. The strength, or contribution, of an element within a given factor is given by its 'factor loading'. A 'factor score' for a given sample can be calculated by determining the sum of these weighted contributions given by each element.

Table 1. Pearson correlation coefficients (x 100) of selected lake sediment data, Strange Lake area. Values between ± 40 are shown by *

	Be	Cd	Ce	Eu	F	Hf	La	Lu	Nb	Pb	Sm	Tb	Th	U	Y	Yb	Zn	Zr	Fe	Mn	LOI
Cd	*																				
Ce	66	*																			
Eu	64	53	78																		
F	42	*	57	57																	
Hf	45	*	*	46	64																
La	62	51	94	78	41	*															
Lu	86	48	78	85	53	54	76														
Nb	54	*	43	46	73	88	*	54													
Pb	79	40	67	71	55	51	61	81	60												
Sm	79	56	88	91	53	44	89	94	45	78											
Tb	87	52	80	88	53	52	79	98	52	83	97										
Th	61	*	63	65	55	47	40	69	57	74	69	69									
U	*	44	*	*	*	*	48	*	*	*	*	*	*								
Y	89	49	80	82	48	48	79	96	50	81	95	99	67	*							
Yb	87	43	76	85	57	61	72	98	62	81	93	98	69	*	97						
Zn	42	66	60	62	*	*	67	53	*	51	62	56	*	53	53	47					
Zr	54	*	46	55	68	96	*	64	91	63	54	62	54	*	59	71	*				
Fe	*	*	*	*	41	*	*	*	*	*	*	*	*	*	*	*	41	*			
Mn	*	*	*	*	51	*	*	*	*	*	*	*	*	*	*	*	*	*	63		
LOI	*	*	*	*	-60	-70	*	*	-73	*	*	*	*	42	*	*	*	-69	*	*	
	Be	Cd	Ce	Eu	F	Hf	La	Lu	Nb	Pb	Sm	Tb	Th	U	Y	Yb	Zn	Zr	Fe	Mn	LOI

Contoured data plots were made of selected data using PLOTGEN, a software package designed by the U.S. Geological Survey, specifically for geochemical data (Evenden and Botbol, 1985). The program was run on the Department of Mines' HP-9000 mini computer. The resulting output is similar to hand-contoured plots, although depending on the selection of various parameters in the package, a certain degree of data smoothing can be obtained. In some instances, this results in individual data points plotting outside their expected contour fields.

From these analyses and data plots it is evident that each area has suites of elements that show enrichments relative to background populations. Histograms and cumulative frequency plots are often effective tools for recognizing and classifying anomalous populations and for selecting threshold values for contoured data maps. A contour map, histogram and cumulative frequency plot for one element from each of the four survey areas, which is representative of that area's rare-metal suite, are presented and discussed below.

Strange Lake Area

Results of the correlation and principal components analyses and of the contoured data maps (McConnell, 1988) indicate that the elements Y, Be, the heavy rare earths, Th, Zr, and Pb and ratios of HREE/LREE in lake sediment reflect the mineralization, and fluoride in water distinguishes the entire peralkaline intrusion. Water samples from the Strange Lake area have less than detection limit concentrations (10 μg/L) of Be, despite their proximity to Be mineralization.

As an illustration of element correlations, a matrix of the Pearson correlation coefficients of the rare-metal suite elements and the environmental variables (Fe, Mn and LOI) are given in Table 1. In general, the REE elements correlate strongly with each other and with Be and Pb. The group Zr, Nb and Hf are strongly correlated with each other, but not with the REE. The elements Zn and U, despite being enriched in the deposit, do not correlate strongly with other elements of the rare-metal suite. The environmental variables, Fe and Mn, correlate moderately with F and Zn, but not with the other members of the suite. LOI has strong negative correlations with F, Nb and Zr, which may explain, in part, why data distribution maps of these elements do not show mineralization dispersion patterns as clearly as do other elements of the suite.

Principal components analysis using a varimax rotation was applied to 36 elements, using logged values. A range of factor models was evaluated, but in each case the first factor was composed of similar elements and included mostly those known to occur in anomalous concentrations in the deposit. The factor loadings (> 0.7) of the elements for the first factor of the three-factor model are given in Table 2. These elements correspond in large part to those associated with the rare-metal deposit. A contour plot of the factor scores of this factor (McConnell, 1988), reveal a very clear down-ice dispersion pattern from the deposit. It is notable that one of the probable ore-elements, niobium, in the Strange Lake deposit, has a rather low-factor loading (0.5). Correspondingly, a contour plot of Nb (McConnell, 1988) gives only a fair dispersion pattern in contrast with those elements with higher loadings.

Table 2. Factor loadings (> 0.70) of mineralization-related factors for a three-factor model of principal components analysis. Lake sediment and water data from Strange Lake and Letitia Lake areas.

Strange Lake	Letitia Lake
Tb (.99)	Y (.93)
Y (.98)	La (.93)
Sm (.97)	Be (.85)
Lu (.97)	Ce (.85)
Yb (.97)	Pb (.84)
La (.88)	Zn (.77)
Th (.87)	Cd (.76)
Be (.87)	
Eu (.87)	
Ce (.86)	
Pb (.86)	
Fw (.81)	

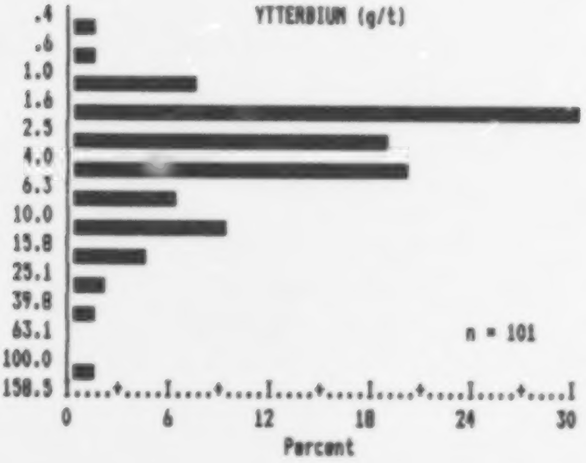

Figure 2. *Histogram of ytterbium distribution in lake sediment, Strange Lake area.*

Ytterbium, a heavy rare-earth element, typifies the distribution in lake sediment of the rare-metal suite associated with the Strange Lake deposit. A histogram (Figure 2) and the cumulative frequency curve (Figure 3) of the log10 Yb distribution, depict the multi-population character of the data. In Figure 2, a population break is apparent between 6.3 and 10 g/t Yb. Moreover, the upper population appears strongly skewed to the right. In the cumulative frequency curve, the population breaks are considerably easier to identify. A pronounced break occurs between 12 and 25 g/t ('A' in Figure 3), which can be seen on the accompanying contour map of Yb (Figure 4) to correspond to sediment values from samples adjacent to, or immediately down-ice from, the main mineralized zone. These are enclosed by the highest contour line of 24.8 g/t. The 6.3 to 10 g/t break seen in the histogram is more clearly defined on the cumulative freqency curve as occurring between 6 and 7 g/t ('B' on Figure 3). This corresponds approximately to the 5 g/t contour line in Figure 4, which encloses samples reflecting dispersion from the entire peralkaline complex. The down-ice dispersion as defined by the contour lines bracketing this population, can be traced for at least 40 km.

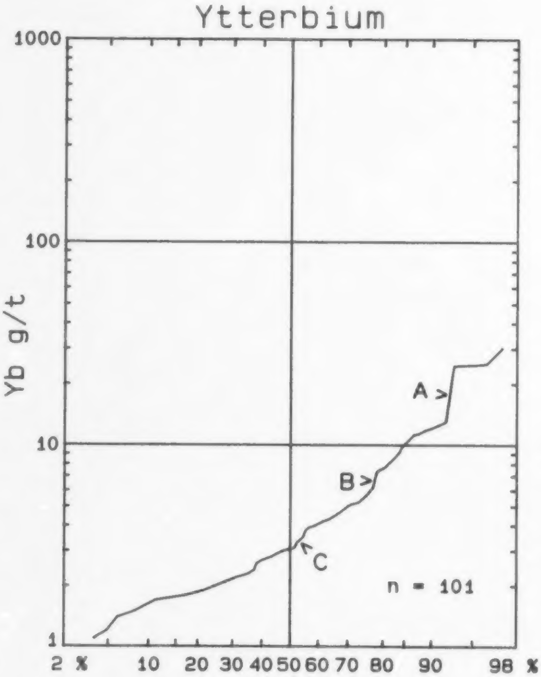

Figure 3. *Cumulative frequency curve of ytterbium distribution in lake sediment, Strange Lake area.*

A third population break, 'C' in Figure 3, occurs at about 3 g/t. Samples in Figure 4, which fall between the 3 and 5 g/t contour lines, are located over and down-ice of the quartz monzonite intrusion. Finally, a fourth population, as defined by the cumulative frequency curve, is composed of those samples falling below 3 g/t. On the contoured data map, these correspond to the area of gneiss terrain to the north of the peralkaline and quartz monzonite dispersion trains, and represent the background population.

Thus, at Strange Lake, four populations (mineralization, peralkaline complex, quartz monzonite and gneiss) can be distinguished by means of a cumulative frequency curve, and these values can then be used to separate samples spatially by provenance on the contoured data map.

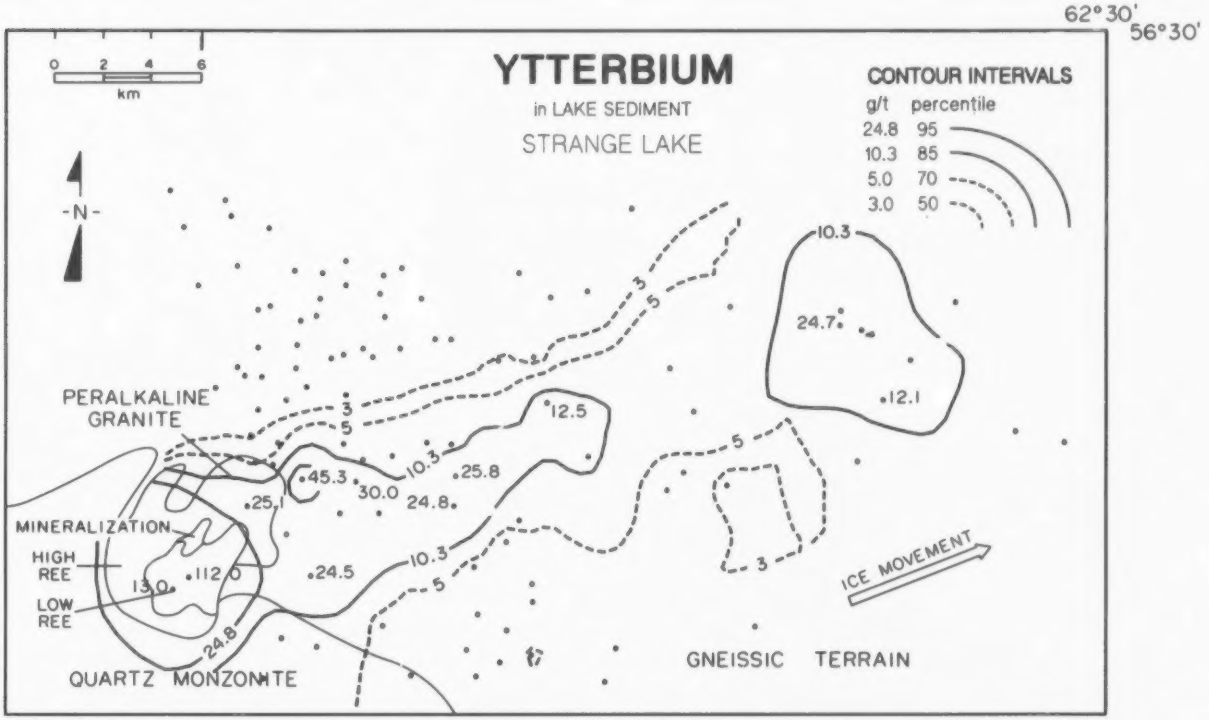

Figure 4. *Contour map of ytterbium in lake sediment, Strange Lake area; (geology from Miller, 1986).*

Letitia Lake Area

The Mann #2 and Two Tom Lake rare-metal occurrences are identified in lake sediment by Zn and Be, and, to a lesser degree, by Nb. Zinc, Y and F in lake waters also highlight the mineralized areas. In addition, unexplored areas are identified, which have Zn, Be, Zr and Nb lake sediment anomalies of equal intensity to those in proximity to mineralization. The Mann #1 occurrence is poorly situated to give rise to a lake sediment anomaly.

Principal components analysis of 24 elements was run and three-, four-, and five-factor models were computed. In each instance, the first factor generated had high loadings on a suite of elements associated with rare-metal mineralization. The elements with the highest loadings (> 0.7) in the first factor of the three-factor model are listed in Table 2. From this, one can see that the elements are similar to the Strange Lake suite shown in Table 2. Note that some of the elements in the Strange Lake factor (i.e., Yb, Tb, Sm, Lu and Eu) were not analyzed in the Letitia survey. A contour plot of this factor (McConnell, 1988) shows that the areas with the highest values are in proximity to the Mann #2 Deposit, the Two Tom Lake occurrence and an area in the south-central part of the survey underlain by rocks of the Red Wine alkaline intrusive suite. This last area is also a beryllium (Be) anomaly and is described in more detail below.

The distribution of Be is discussed as representative of the rare-metal suite in the Letitia Lake area. A histogram of the Be distribution (Figure 5) and a cumulative frequency curve (Figure 6) are presented. In both figures, population breaks are discernible but are not as obvious as with the corresponding figures for Yb at Strange Lake. The histogram appears generally log-normal with a small subpopulation above 63 g/t. The cumulative frequency curve indicates that an inflection point occurs somewhere between 40 and 70 g/t ('A' in Figure 6) separating the highest population from the rest of the data. This break is represented on the contoured data map of Be (Figure 7) by the 46 g/t contour line. Sample sites within this contour include those in the vicinities of the Mann #2 deposit and the Two Tom Lake occurrence. In addition, a large anomaly is outlined in the south of the survey area centred over a portion of the Red Wine alkaline intrusive suite. A fourth anomaly, represented by a single sample (69 g/t), is located 9 km southwest of Mann #2.

Two inflection points on the cumulative frequency curve are denoted by 'B' and 'C' in Figure 6. The 'C' inflection at about 23 g/t corresponds to the second highest contour line in Figure 7. Most of the samples enclosed by this contour are from an area extending up-ice from the Two Tom Lake occurrence and from an area surrounding the southern anomaly. In the first area, these samples suggest that either there is an area of bedrock enrichment surrounding the Two

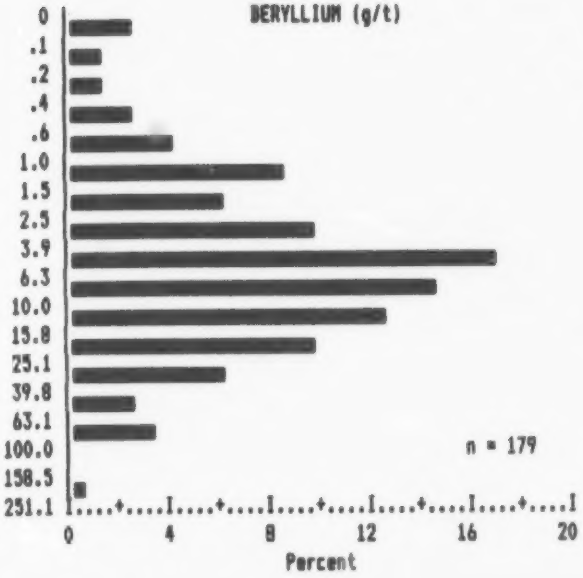

Figure 5. *Histogram of beryllium distribution in lake sediment, Letitia Lake area.*

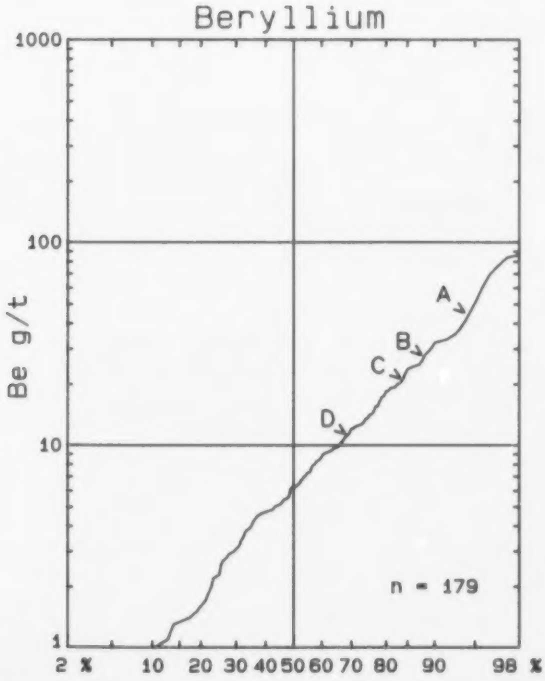

Figure 6. *Cumulative frequency curve of beryllium distribution in lake sediment, Letitia Lake area.*

Tom mineralization or that there is additional mineralization not yet recognized. In the second area, the samples extend the size of the as-yet unexplained anomaly.

A final minor inflection point, 'D' at about 11 to 12 g/t Be, corresponds to the lowest contour line in Figure 7. This contour line roughly outlines the extent of Letitia Lake Group rocks in the north and also expands the southern anomaly.

Michikamats Area

There are anomalies present in F, Ce and La similar in magnitude to those associated with mineralization at Strange Lake and Letitia Lake. However, the concentrations of the elements Y, Nb and Be are substantially lower than in the mineralized areas. No HREE data are available for this area, but since the chemical behaviour of HREE is similar to that of Y, levels are not likely to be high. The dispersion pattern of F in water (Fw) is similar to those of Be, Y and La and is discussed here as a representative element.

The histogram of Fw (Figure 8) has an irregular shape. The cumulative frequency curve (Figure 9) shows several distinct breaks separating at least 5 component populations. However, the contour levels chosen for the map (Figure 10), only correspond in part to these breaks. The highest break ('A' at about 550 to 580 μg/L) is a minor one and does correspond to the highest contour level in Figure 10. This encloses only a single sample, 1867 μg/L. The other six samples with values > 580 μg/L are surrounded by samples with somewhat lower Fw contents, whose influence prevents the higher samples from being separately contoured.

A second break, 'B', occurs at about 350 to 400 μg/L. This is approximately equivalent to the second contour line at 300 μg/L, which delineates two centres of high Fw concentrations that are also reflected by Be, Y and La (McConnell, 1988). The lowest contour (224 μg/L) does not correspond to an inflection point but, rather, occurs midway between the two points 'B' and 'C'. The sample sites enclosed by it are associated with the northern anomaly. A break at 60 μg/L, 'D', separates the lowest 15 to 20 percent of the data that are not identified on the contour map.

13L Area

Results from the 21 samples in this small area indicate an area of enrichment in Y and F similar in magnitude to element levels associated with mineralization at Strange Lake and Letitia Lake. Levels of Be and Ce are also moderately high, whereas those of Zr and Nb are not.

A bimodal distribution of Y is apparent in the histogram (Figure 11) and the cumulative frequency curve (Figure 12) with a break ('C') occurring between 25 and 60 g/t. Lesser inflection points occur at 170 g/t ('B') and at 220 g/t ('A').

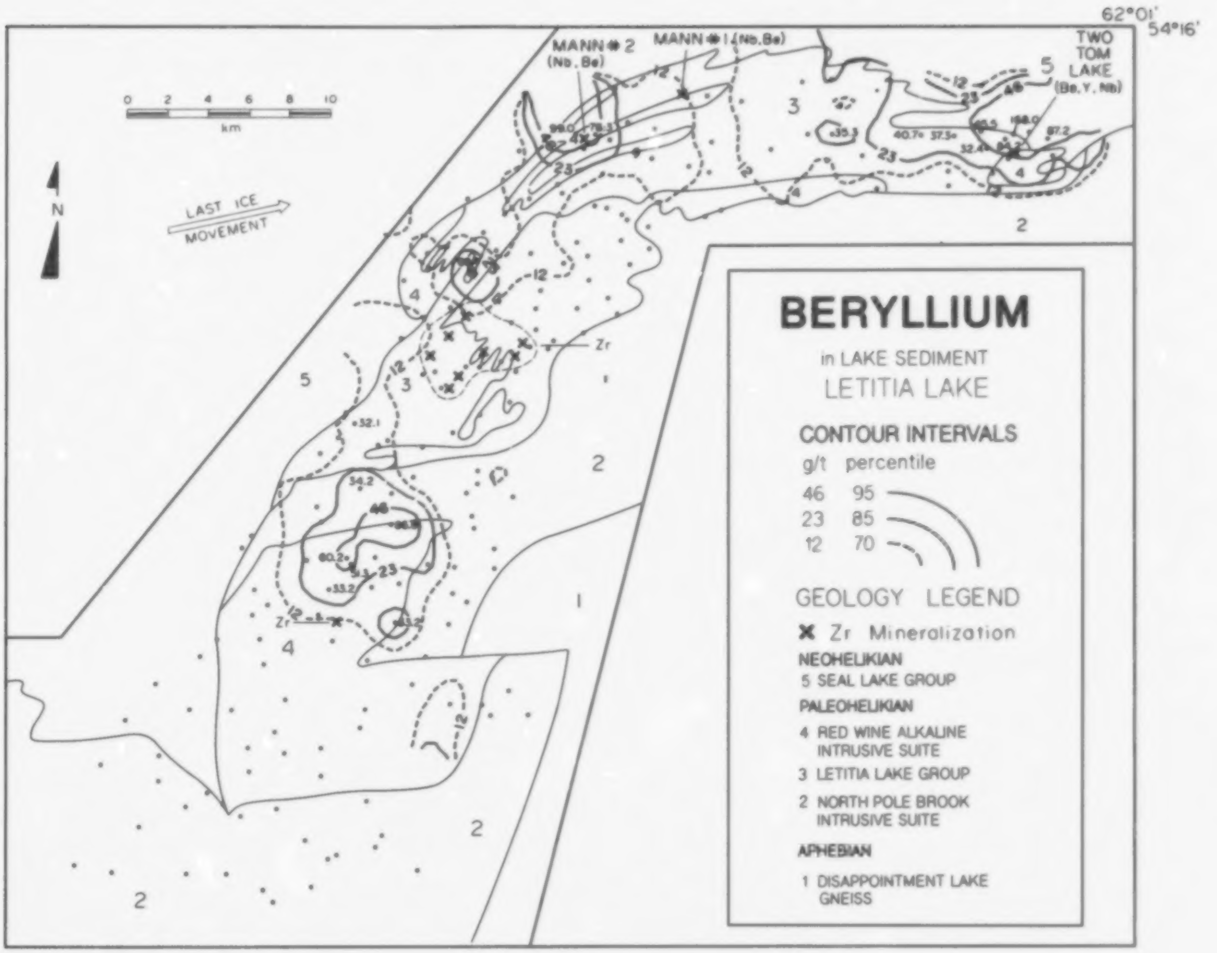
62°01'
54°16'
MANN # 2 (Nb, Be)
MANN # 1 (Nb, Be)
TWO TOM LAKE (Be, Y, Nb)
0 2 4 6 8 10
km
N
LAST ICE MOVEMENT
Zr
BERYLLIUM
in LAKE SEDIMENT
LETITIA LAKE
CONTOUR INTERVALS
g/t percentile
46 95
23 85
12 70
GEOLOGY LEGEND
✖ Zr Mineralization
NEOHELIKIAN
5 SEAL LAKE GROUP
PALEOHELIKIAN
4 RED WINE ALKALINE INTRUSIVE SUITE
3 LETITIA LAKE GROUP
2 NORTH POLE BROOK INTRUSIVE SUITE
APHEBIAN
1 DISAPPOINTMENT LAKE GNEISS

Figure 7. *Contour map of beryllium in lake sediment, Letitia Lake area (geology after Thomas et al., 1983 and Thomas and Hibbs, 1983).*

The upper point is similar to the upper contour line (Figure 13) of 210 g/t. A pair of sample sites are grouped within this contour. The next lower contour, 189 g/t, includes only 2 samples and 1 sample of 191 g/t falls outside the line. The delineation of the dispersion pattern would be improved by replacing the 189 g/t contour with 170 g/t. Similarly, replacing the 140 g/t line with a 40 g/t line might better resolve the pattern in terms of underlying bedrock. Regardless, there is distinct multi-sample anomaly centred in the southeast portion of the survey area.

CONCLUSIONS

The following conclusions can be drawn from this study:

1) Many elements in the rare-metal suite are present in samples of lake sediment and water from mineralized areas in concentrations that are considerably higher than in samples from non-mineralized areas. Thus, these media provide a potent prospecting method. In sediment, these elements include Be, F, Hf, Nb, Pb, REE, Th, Y, Zn and Zr. In water, they comprise F, Y and Zn. The ability of an element to disperse into lake sediment depends partly upon the stability of the source minerals in the weathering environment.

2) Two methods of analysis, ICP and direct neutron activation, are particularly valuable in providing these data.

3) Principal components analysis offers a practical technique for recognizing element associations diagnostic of rare-metal mineralization. This is particularly useful in areas where knowledge of the chemical character of the target mineralization is lacking.

4) Environmental factors encountered within the lake such as organic content (LOI), pH of water, and adsorption by Fe and Mn oxides, do not appear to have a large influence on the distribution of REE in lake sediment. The distributions of F, Hf, Nb and Zr, however, have moderate to strong

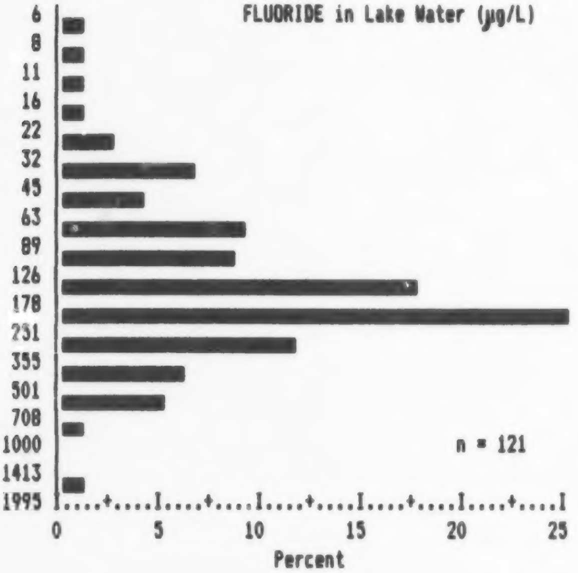

Figure 8. *Histogram of fluoride distribution in lake water, Michikamats area.*

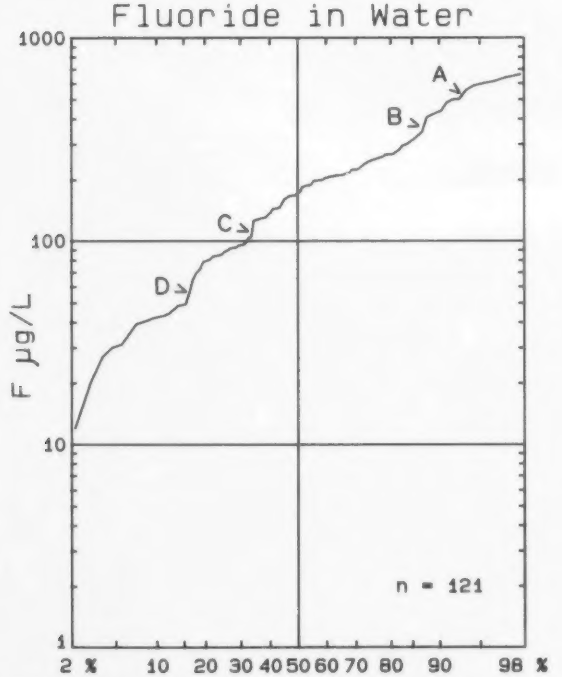

Figure 9. *Cumulative frequency curve of fluoride distribution in lake water, Michikamats area.*

negative correlations with LOI in most areas. The interpretation of dispersion patterns of such elements may be enhanced by the use of regression analysis to calculate residual values.

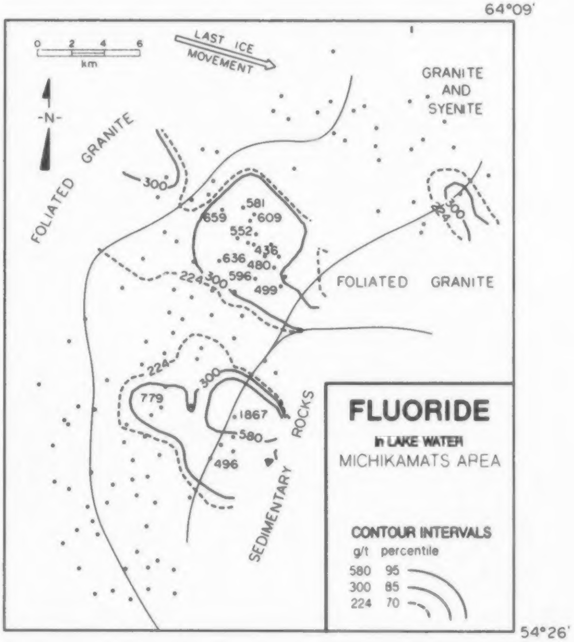
64°09'
54°26'
0
2
4
6
km
-N-
LAST ICE MOVEMENT
FOLIATED GRANITE
GRANITE AND SYENITE
FOLIATED GRANITE
SEDIMENTARY ROCKS
300
581
659
609
552
436
636
480
596
499
224
300
300
224
224
300
779
1867
580
496
FLUORIDE
in LAKE WATER
MICHIKAMATS AREA
CONTOUR INTERVALS
g/t percentile
580 95
300 85
224 70

Figure 10. *Contour map of fluoride in lake water, Michikamats area; (geology from Emslie, 1963).*

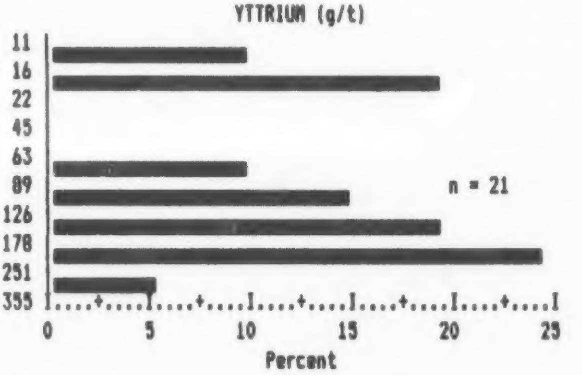

Figure 11. *Histogram of yttrium distribution in lake sediment, 13L area.*

5) The shape and areal extent of geochemical dispersion fans vary broadly and are products of the erosional and depositional regimes active during glaciation. In the Strange Lake area, these processes produced a geochemical fan identifiable for up to 40 km down-ice from the deposit. By contrast, glacial smearing is much less evident in the other three areas where maximum anomaly displacement from source is on the order of 2 to 4 km.

6) Several unexplained anomalies are present in areas that have received little or no exploration in the Letitia Lake,

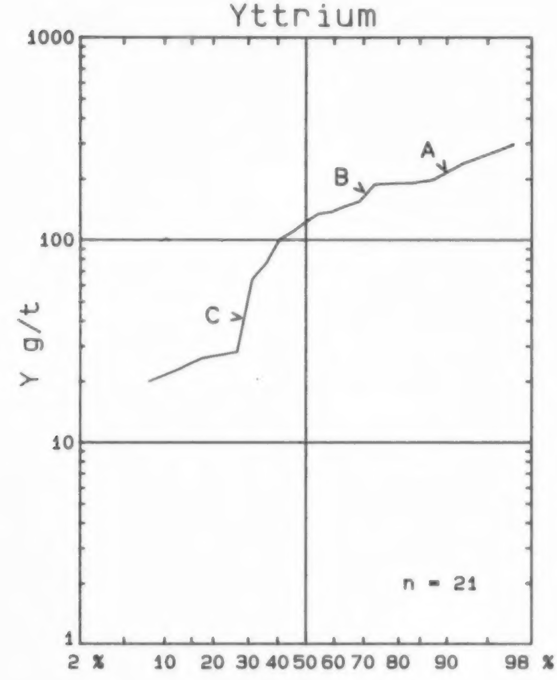

Figure 12. *Cumulative frequency curve of yttrium distribution in lake sediment, 13L area.*

Michikamats and 13L areas. These targets are potential sites of undiscovered rare-metal mineralization.

ACKNOWLEDGMENTS

This report is based on two seasons of field work, to which many individuals contributed. In particular, I would like to thank Randy Miller and Martin Batterson for their support and advice in the Letitia and Strange lake areas. Lisa Newman prepared some of the figures and did much of the early data editing. Peter Davenport is thanked for his critical reading and suggestions, which improved the final version of the manuscript.

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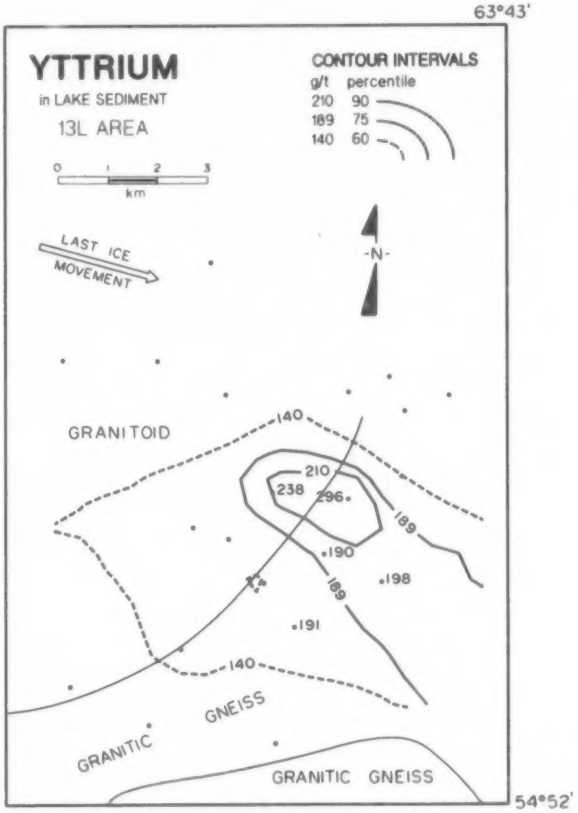
63°43'
YTTRIUM
in LAKE SEDIMENT
13L AREA
CONTOUR INTERVALS
g/t percentile
210 90
189 75
140 60
0 1 2 3
km
LAST ICE MOVEMENT
-N-
GRANITOID
140
210
238
296
189
190
198
189
191
140
GNEISS
GRANITIC
GRANITIC GNEISS
54°52'

Figure 13. *Contour map of yttrium in lake sediment, 13L area; (geology from Greene, 1970).*

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NEWFOUNDLAND AEROMAGNETIC TOTAL-FIELD, GRADIOMETER AND VLF-EM SURVEY, 1986/87

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Geophysics Division
Geological Survey of Canada, Ottawa

ABSTRACT

An aeromagnetic total-field, gradiometer and VLF survey was carried out over part of the Dunnage Zone in central Newfoundland. This part of the Dunnage Zone is geologically complex and well suited to aeromagnetic mapping. Aerodat Limited of Toronto carried out the contract in 1986–87 using a rotary wing aircraft.

Results from the survey are available at two map scales: 1:25,000 aeromagnetic total field and gradiometer contour maps, and 1:50,000 magnetic anomaly and gradiometer colour interval maps. The VLF-EM total field and quadrature profiles are printed on the back of the 1:50,000 maps.

The aeromagnetic results substantiate and expand upon the known geology in this part of the Dunnage Zone. In particular, the gradiometer colour maps show clearly defined intrusive features and major transcurrent faults. A series of folded sills, as well as a thrust sheet, can be identified in the middle and northern regions of the survey area.

INTRODUCTION

Under the Canada–Newfoundland Mineral Development Agreement, an aeromagnetic total-field, gradiometer and VLF-EM survey was carried out over an area within the Dunnage Zone of central Newfoundland. A sketch of the survey location is shown in Figure 1. The complexity of the basement in this area is such that its structural and compositional changes have a definite magnetic signature, making indirect geological mapping through aeromagnetics very successful.

The contract to carry out the field survey was awarded to Aerodat Limited of Toronto, and the survey was flown between November 1986 and August 1987, using a rotary wing aircraft. Two Cesium vapour magnetometers separated by 3 m, each with a resolution of 0.002 nanoteslas, measured the gradient field. VLF-EM total field and vertical quadrature components were recorded with a Herz Totem-2A receiver, using the Annapolis, Maryland transmitter signal. A mean elevation clearance of 150 m was maintained with a flight line separation of 300 m and a control line separation of 5 km.

Survey results are available at two map scales: 1:25,000 aeromagnetic total-field and gradiometer contour maps and 1:50,000 magnetic anomaly and gradiometer colour interval maps. The VLF-EM total-field and quadrature profiles are printed on the back of the magnetic anomaly and gradiometer colour interval maps, respectively. This presentation provides a means to view the VLF-EM data and the magnetics with the use of a light table. In addition, the digital data can be purchased for further processing and interpretation from the Geophysical Data Centre, Geological Survey of Canada, 1 Observatory Crescent, Ottawa, Ontario, K1A 0Y3, (613) 992-6438.

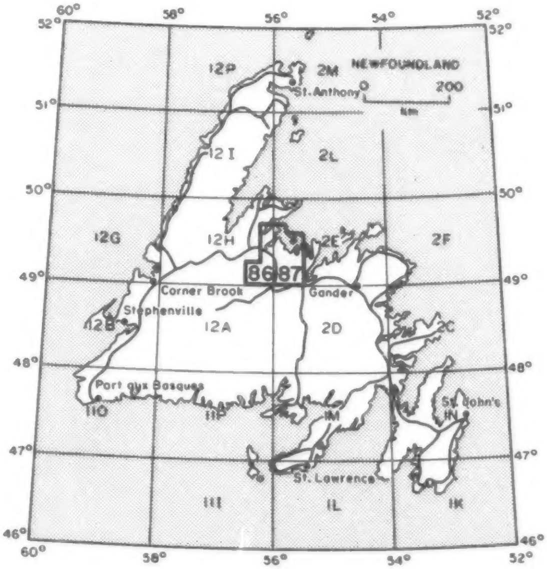

Figure 1. *High resolution aeromagnetic total-field, gradiometer and VLF-EM survey 1986-87, central Newfoundland.*

Plates 1 and 2 are photo reductions of the magnetic anomaly and gradiometer maps, respectively. Superimposed on each is an overlay emphasizing the major magnetic markers.

This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

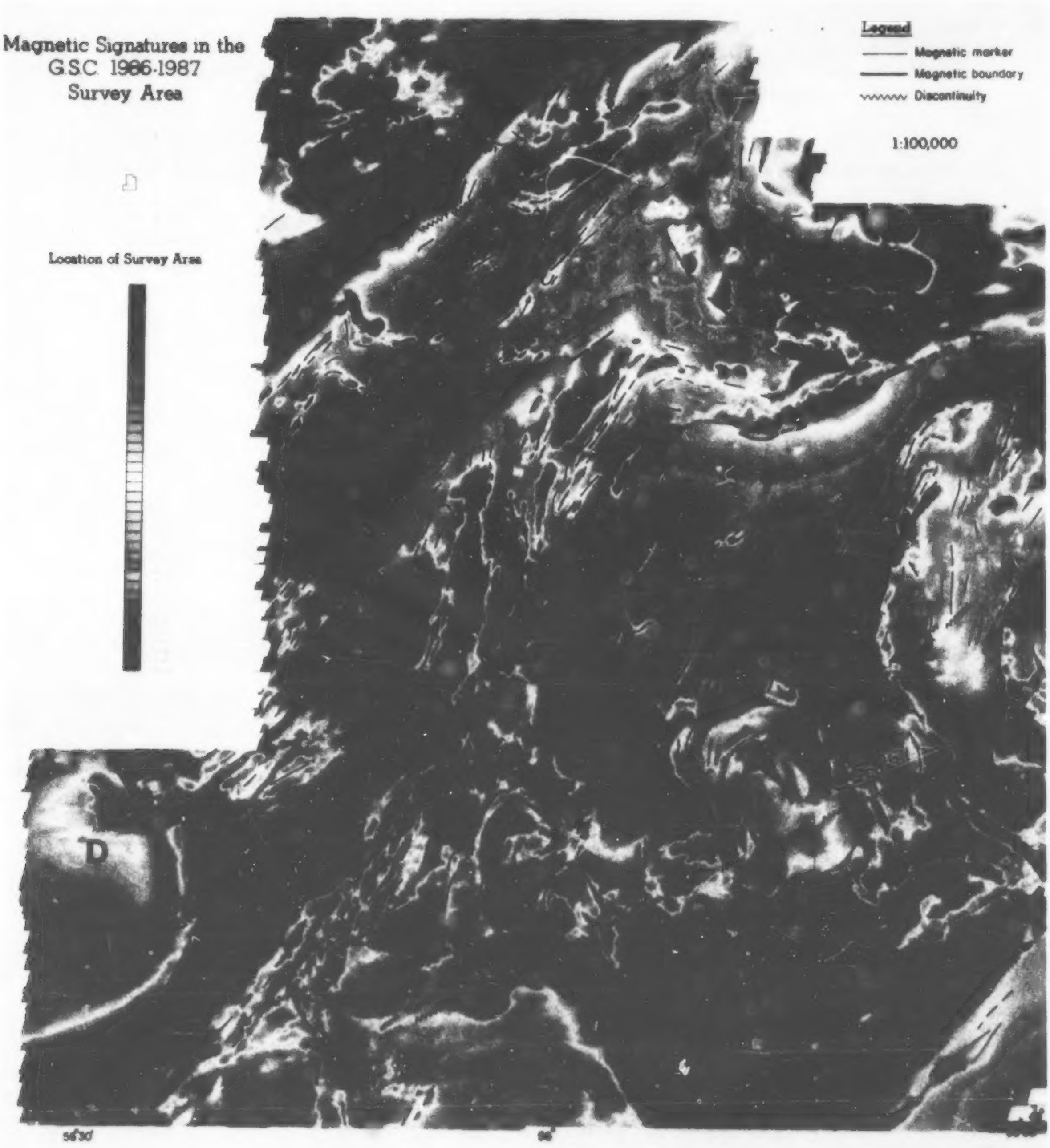
Magnetic Signatures in the G.S.C. 1986-1987 Survey Area
Location of Survey Area
Legend
Magnetic marker
Magnetic boundary
Discontinuity
1:100,000
D
56°30'
56°

Plate 1. *Magnetic anomaly map with major magnetic markers outlined, central Newfoundland.*

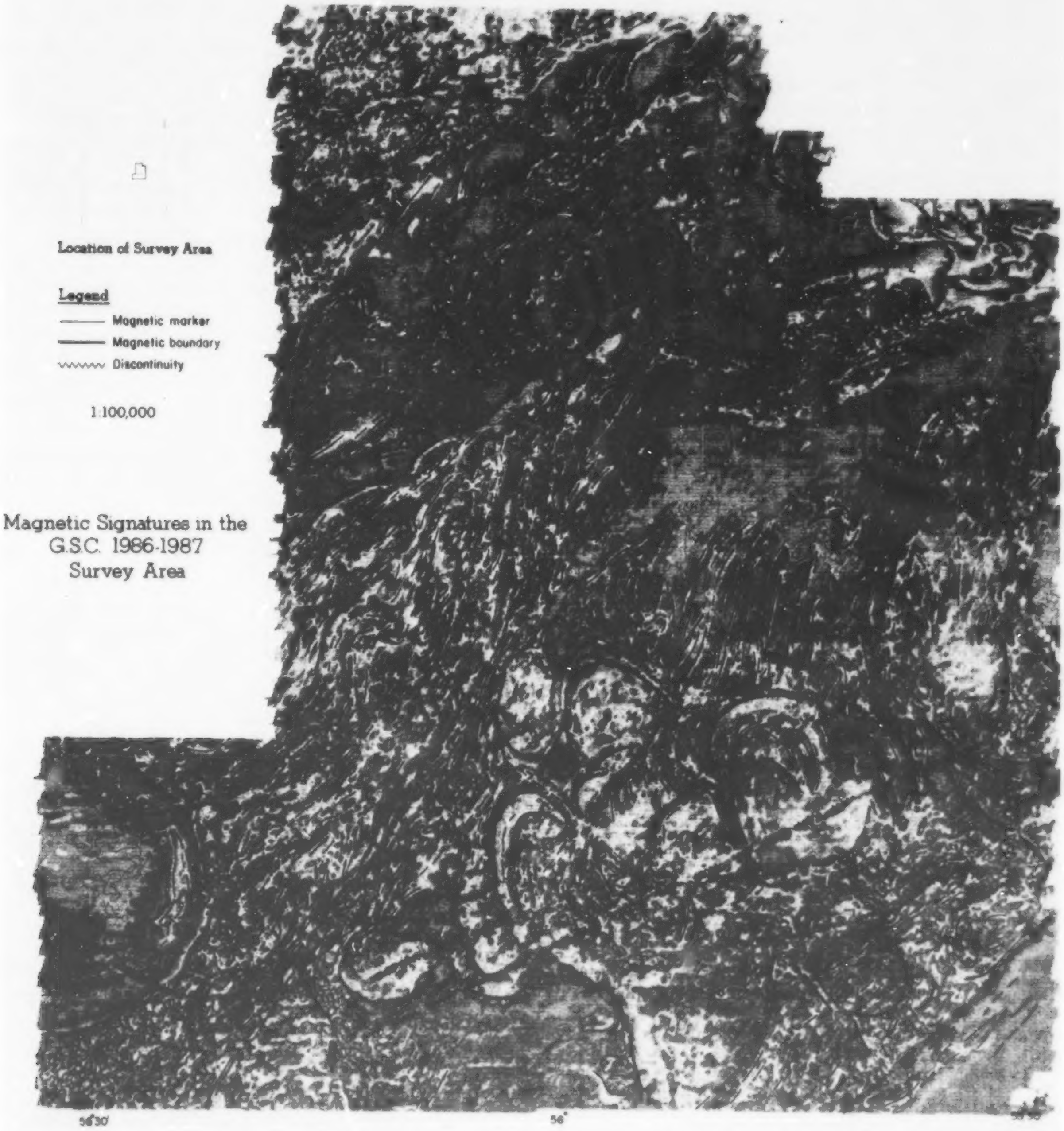
Location of Survey Area
Legend
Magnetic marker
Magnetic boundary
Discontinuity
1:100,000
Magnetic Signatures in the G.S.C. 1986-1987 Survey Area
56°30'
56°
55°30'

Plate 2. *Vertical gradient map with major magnetic markers outlined, central Newfoundland.*

RESULTS

In this central Newfoundland survey area, the magnetic signature confirms mapped geology (Dean, 1977) and provides additional details. A number of structural features stand out clearly on the gradiometer colour maps. Several phases of plutonic intrusion have occurred in the southern part of the survey and these are transected by more recent northeast-striking faults. A folded swarm of sills intrudes farther to the north. Near the top of the survey area, there is evidence of a thrust sheet. Although all of these features have been previously identified, the specific information on each is enhanced by the geophysical results.

The igneous intrusions in the south ('A', Plate 1), are Devonian (Dean, 1977), and are graphically outlined by the magnetics, but only to a lesser degree by the VLF-EM. They are identified by a horseshoe-shaped pattern, which characterizes the magnetic contrast along their perimeter. The Hodges Hill, Topsails, and Dawes Pond granites, as well as the Twin Lakes Diorite Complex contacts, are particularly well defined. Within certain intrusive units, there is a secondary magnetic signature indicating either magmatic differentiation or possibly, progressive stages of intrusion.

The gradiometer trace is particularly effective in identifying magnetic contacts, as is the case with the Devonian igneous intrusions described earlier. The zero gradiometer contour tends to lie directly above a magnetic contact in northern latitudes. For Newfoundland, this contour will be slightly offset due to the inclination of the earth's field, the effect of remanent magnetization and the interference from adjacent sources. VLF-EM anomalies are associated with both contrasts in conductivity between rock units, and cultural features. This survey area is dotted with bodies of water, swamps, creeks, etc., which complicate the interpretation of the VLF-EM signature.

In the north ('B', Plate 1), the ophiolitic rocks of the Lushs Bight Group are thrust faulted against the Silurian Springdale and Ordovician Roberts Arm groups (Kean, 1984). The magnetics trace this fault (Lobster Cove Fault) very clearly, where the basaltic pillow lavas are thrust against the Roberts Arm Group to the south. The sharpness of this delineation decreases westward.

Middle and Lower Ordovician diabase and gabbro sills intrude the Wild Bight Group southwest of Badger Bay ('C', Plate 1). The sills appear magnetically as a cluster of several, parallel to sub-parallel, folded, elongated anomalies. The contrast in magnetic response between these units is such that folding patterns can be readily traced. The vertical gradient indicates that the number and extent of these folded, north–northeasterly trending units is far greater than currently identified from field mapping.

The Topsails granite ('D', Plate 1) is located along the most westerly edge of the survey area. Immediately to the north, are the volcanics of the Silurian Springdale Group, which are alternately acidic and mafic units. The difference in magnetic signature between these alternating units, helps to define the Burnt Berry syncline, which trends to the northeast.

A post-Devonian period of major faulting is strongly evidenced in the magnetic data and these east–northeasterly faults ('E', Plate 1), such as Northern Arm, Long Pond, and Four Mile faults show up clearly (particularly where magnetic features are intersected at angles near 90°). The faults are extensive, readily evident and in several cases, clearly indicate transverse displacement. Several of these previously mapped faults can be extended or modified on the basis of this new information.

These structural and lithological features that have been highlighted are only those that are the most apparent on the 1:50,000 colour maps. Further qualitative details, in addition to a great deal of quantitative information, (e.g., depth to source rock, susceptibility, shape of causative body) can be derived from the 1:25,000 contour maps and the profile data.

Funds for this project were provided by the Geological Survey of Canada under the Canada–Newfoundland Mineral Development Agreement, 1984-1989.

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ENHANCEMENT AND INTERPRETATION OF REGIONAL GEOPHYSICAL DATASETS IN CENTRAL NEWFOUNDLAND

Gerry Kilfoil
Geochemistry and Geophysics Section

ABSTRACT

High resolution aeromagnetic data for an area south of Buchans in the Central Volcanic Belt was digitally filtered and displayed on an image analysis system. Several lithological units are readily identified by their magnetic signature on images of the data, shaded using a false sun inclined 45° from the northeast.

Structural lineations trending north-northeast locally interrupt the northeast regional trend of magnetic units, and are interpreted to be an expression of flexure at the latter stages of sinistral strike-slip movement along the major northeast-oriented regional faults. A series of northwest-oriented late mega-fractures', apparently transecting all earlier structures, are traced across the entire breadth of the Central Volcanic Belt. These mega-fractures exhibit only minor offsets on steeply dipping, northeast-oriented magnetic units identified in the data, and thus are interpreted as expressing vertical movement on structural units that cannot be easily identified by surface mapping. These late meta-fractures are most probably an expression of isostatic readjustment or tension.

The structural features described herein are compared and related to known occurrences of massive sulphide and precious-metal mineralization in the Central Volcanic Belt.

INTRODUCTION

During 1987 and 1988, a project was undertaken to manipulate and display available high resolution airborne magnetic, vertical gradiometric and VLF-EM datasets collected throughout the belt of volcanic and sedimentary rocks of the Victoria Lake Group near Buchans (areas 1 to 4 on Figure 1) by the Geological Survey of Canada. The results of the survey in the Notre Dame Bay area (area 5 on Figure 1), have only recently been released (Tod and Ready, *this volume*). Digital data from the 1985 survey, areas 1 to 3, were selected as the focus of this investigation for reasons of recent or concurrent mapping, the variety of geological units represented and its mineral potential. A series of total-field magnetic and gradiometric colour maps of the data from this GSC survey have been published at 1:50,000 scale (Geological Survey of Canada, 1985a,b,c,d,e). The purpose of this study is to digitally filter and combine these geophysical datasets in such a manner as to gain additional information that would reflect the underlying geology and structure, not readily apparent on the surface.

METHODOLOGY

Total-field aeromagnetic and vertical magnetic gradiometer datasets were pre-processed using the Newfoundland Department of Mines' mini-computer system, transferred to digital tape and loaded onto the Dipix Aries III image analysis system at NORDCO Limited for display and hardcopy output. Digital processing consisted of IGRF

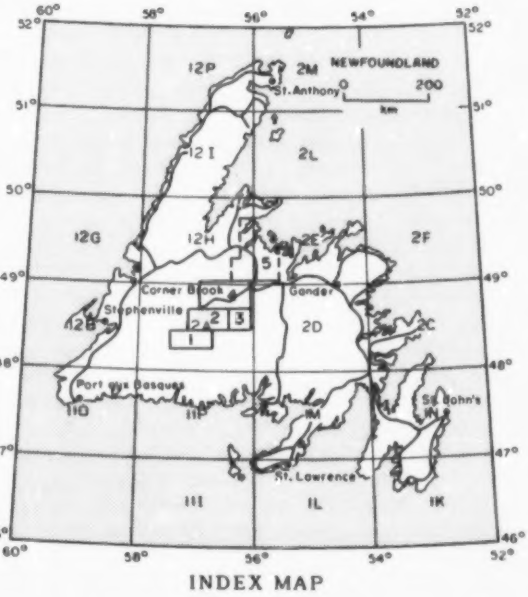


Figure 1. *Location map in central Newfoundland of Geological Survey of Canada high-resolution airborne total-field magnetic and gradiometer surveys carried out under the Canada–Newfoundland Mineral Development Agreement, 1984-89. Areas 1-3: Lower Victoria Lake Group–Noel Paul's Line; area 4: Buchans–Badger; area 5: Notre Dame Bay.*

This project is a contribution to the Canada–Newfoundland Mineral Development Agreement, 1984–1989

removal from the total-field data, variable false sun-angle shaded relief, and calculated first and second vertical derivatives. The digitally filtered total-field data images included as Figures 2a and 3a are mosaics of photographed images. Although a series of colour images were produced, only the grey-tone versions are included in this publication.

Directional filters of varying angles of false sun illumination were applied to the total-field magnetic (IGRF removed) and vertical gradient datasets to optimally enhance their subtle features (Teskey and Broome, 1984), as well as features that have a large relative magnetic relief. As a result, some information regarding the total-field strength of the local magnetic field is sacrificed in the filtering process. The various calculated gradient maps (first and second derivatives), and the gradiometric data themselves, were shown to be valuable interpretive tools in tracing the boundaries of local bedrock lithologies, particularly in areas of overburden. A series of shaded-relief aeromagnetic maps have proven to be of most value as an interpretational aid to the geologist or geophysicist.

First-order interpretation of characteristic magnetic signatures, such as those expressed by volcanogenic sediments, nearly homogeneous plutonic rocks and large-scale flexures, correspond to those identified on the surface. In addition, several small-scale and subtler features are enhanced by specific illumination directions. Among these are foliation orientations within plutonic bodies and a series of large-scale, late-stage, northwest-oriented fractures not readily detected on the surface. The latter features are of particular interest because of their relationships to gold and massive sulphide mineralization within the Victoria Lake Group rocks.

RESULTS

Areas 1 and 2

The grey-tone photo-mosaic of Figure 2a shows the image, which results from directional filtering applied to aeromagnetic data for areas 1 and 2 (Figure 1), using a false sun inclination of 45° and declination of 45°. Figure 2b shows the boundaries of major geological units (dotted line) compiled from quoted sources, and the geophysically identified geological unit boundaries and structural lineations (solid and dashed lines, respectively) identified in this study. In addition to 'real' magnetic features enhanced on the image of Figure 2a, the presence of any local data-levelling errors resulting from corrections applied between adjacent east–west flight lines are exaggerated by shading from this oblique angle. Leveling errors, or flight-line noise, gives the image a corrugated appearance that is most pronounced in Figure 2a, along latitude 48°30'N. The regular 'bubbly' appearance of several of the northeast-trending magnetic features throughout the image is an artifact of their oblique trends relative to the east-west direction of both flight lines and the orientation of gridding.

The variations in trend of magnetic units from northeast to north-northeast are obvious in Figure 2a, as traced by the solid and dashed lines in Figure 2b. Magnetic anomalies associated with units in the region commonly have an asymmetrical, elongated lobate shape resembling a magnetic hysteresis loop. This diamond shape is an expression, on a local scale, of the regional flexure resulting from changing stresses during deformation in a sinistral strike-slip regime (Kean and Evans, 1988).

In the upper central part of Figure 2a, the belt of magnetic units trending to the northeast are the Victoria Lake Group mafic to felsic volcanic and sedimentary rocks ('A' in Figure 2a) of the Central Mobile Belt. Linears defined by variations in trends of these units are illustrated by dashed lines on Figure 2b. The Victoria Lake Group units pinch out in a major flexure located near the southwest corner of this image (just north of 'C', Figure 2a).

The large, foliated granite (informally named the Snowshoe Pond granite), located at 'B' in the south-central region of Figure 2a, shows a distinct non-magnetic character and is evident here as a lobate area of low relief. A large linear magnetic anomaly forms the southwestern boundary of this lobate pluton and separates it from the Buck Lake granite pluton of the North Bay Granite ('C' in Figure 2a). Aside from the flight line interference present in the southwestern portion of Figure 2a, the Buck Lake granite shows a mottled appearance on the shaded relief image relative to that of the Snowshoe Pond granite, indicating its greater variability in magnetic content. The nature of this seemingly sharp contact between the two plutons is unknown, as this area is characterized by swamp or bog and no significant lithological difference has been reported (Kean, 1977a). As a result, the two plutons have not been previously distinguished as distinct map units (Kean, 1977a, 1982). A gabbro unit is mapped just north of the Buck Lake granite (Kean, 1982) ('D' in Figure 2a) where the northwest oriented magnetic division between the two plutons obliquely intersects the Victoria Lake Group sediments to the northwest. The continuity in magnetic expression of the boundary between the two plutons southeastward from the gabbro suggests that the gabbroic or mafic dyke rocks may be found along the contact between the two plutons. Alternatively, the anomaly may arise from migmatized Spruce Brook Formation sediments located to the southeast (Colman-Sadd, 1988). A magnetic 'striping' can be observed within the Buck Lake granite boundary of Figure 2a, and running parallel to its contact with the Snowshoe Pond granite. This may indicate that the Buck Lake granite and bordering country rocks have been variably altered along this contact during its emplacement.

The southeastern area on Figure 2a (at 'E') is underlain by granites and metasediments of the Meelpaeg Subzone of the Gander Zone (Williams, *et al.*, 1988; Colman-Sadd, 1988). The high-grade, quartz-rich metasediments equated to the Spruce Brook Formation (Colman-Sadd, 1988) show a variable magnetic pattern and a northeast oriented layering that is less pronounced than that of the Victoria Lake Group. The magnetic expression of the metasediments is further complicated by the presence of at least two varieties of Silurian to Devonian intrusive rocks; these are foliated to locally mylonitic or polydeformed, syntectonic granites (Snowshoe Pond granite) and undeformed to slightly foliated, probable posttectonic units (Kean, 1982; Colman-Sadd, 1988).

The Valentine Lake quartz monzonite, located north of the Snowshoe Pond granite, can be identified on Figure 2a ('F') by the curvilinear magnetic anomaly trends that form its outline. The variable magnetic field strengths within the pluton are due to its phases, which vary from granitic to gabbroic in composition. A wedge of metasediments ascribed to the Meelpaeg Subzone separate the Valentine Lake quartz monzonite from the Snowshoe Pond granite to the south. A lens of Rodeross Lake gabbro is identified just to the east ('G' in Figure 2a) by its variable appearance, in sharp contrast to the magnetic signature of the Snowshoe Pond granite that nearly encompasses it.

The Rogerson Lake Conglomerate is characterized on Figure 2a ('H') by a low magnitude, smooth magnetic linearity, the boundaries of which are defined by magnetic anomalies. Though the conglomerate narrows to 200 m along the southeastern border of the Valentine Lake quartz monzonite (Kean, 1982), the belt continues uninterrupted from beyond Noel Paul's Brook map area east of Figure 2a to the shores of Victoria Lake in the southwest. This unit is thought to mark the presence of a major tectonic break (Kean and Evans, 1988). The Rogerson Lake Conglomerate separates members of the Victoria Lake Group to the northwest from similar volcanic-rich units to the southeast, which are considered (Colman-Sadd, 1988) as Victoria Lake Group equivalents.

The mottled magnetic pattern that may be observed in the northwest corner of Figure 2a ('I') arises from variably magnetic gabbros and granites within the Topsails Intrusive Suite, which overthrust the Victoria Lake Group (Kean, 1978, 1979, 1982). The Skidder basalt can be identified in Figure 2a (at 'J') by its characteristic magnetic pattern. The large area of magnetic low relief, that flanks the Topsails Igneous Suite to the southeast ('K' in Figure 2a) corresponds to the location of a small Carboniferous subbasin (Kean, 1979) filled with magnetically transparent sediment, similar in character to that of Carboniferous subbasins found elsewhere in Newfoundland (*cf.* Kilfoil, 1988). A linear wedge of Harbour Round Formation volcaniclastic sediments (Kean, 1977b) is situated at the southeastern margin of the Carboniferous sedimentary subbasin ('L' in Figure 2a). The elevated magnetic total-field values observed for this unit in Figure 2a are due to mafic lava units within the siltstones, cherts and breccias.

The 45° declination angle of illumination delineates a series of large-scale, late-stage, northeast-oriented lineaments (as indicated by arrows on Figure 2a) corresponding to gradiometer anomalies of Kean and Evans (Figure 4, 1988). Locations of known gold and/or massive sulphide occurrences, indicated by dots on Figure 2b, have been included for the purpose of correlation with identified structural lineations. The subtle northwest-oriented features are of interest for possibly providing a conduit system along which hydrothermal alteration and mineralization within rocks of the Victoria Lake Group occurred.

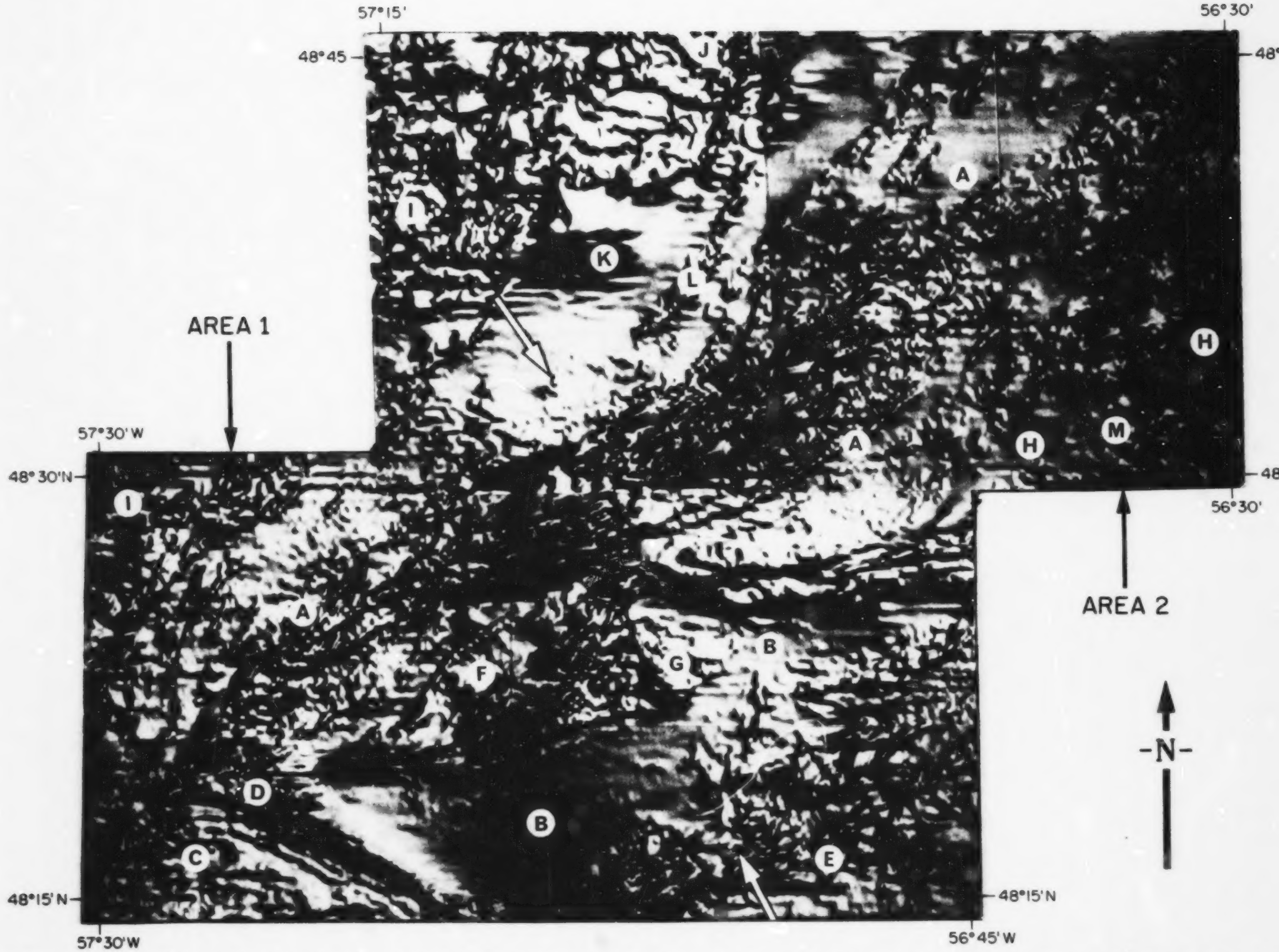
Although not displayed here, the flight line 'noise' that pervades the image is appreciably diminished on images generated by shading from 90° declination as this illumination direction is parallel to that of flight lines. In addition, the subtle expression of a north-northeast oriented foliation within the Snowshoe Pond granite is notably enhanced on the image shaded from the east.

Area 3

Figure 3a contains an image of the gridded aeromagnetic data for Noel Paul's Brook (NTS 12A/9, Area 3 of Figure 1) enhanced by the same false sun direction as that of Figure 2a. Overlap of the western border of Figure 3a with the eastern border of Figure 2a allows several features to be matched. Figure 3b shows both the established boundaries of the major geological units (dotted line) within the study area and the geophysically identified geological unit boundaries and structural lineations (solid and dashed lines, respectively) identified in this study.

The most prominent feature on Figure 3a is the smooth, low-relief region underlain by the non-magnetic Overflow Pond Granite (at 'A') (Kean and Jayasinghe, 1980), which occupies a large area in the east-central portion of the map area. This coarse grained, two mica granite exhibits a consistently low magnetic field in which only the minor effects of the flight lines are apparent when digitally enhanced. The geophysically interpreted southern boundary of the granite is situated south of that mapped by Jayasinghe (1978). However, geological mapping in this portion of the map area is based on sparse outcrops. The two circular magnetic anomalies located within the area occupied by granite, just northeast of 'A' on Figure 3a, are likely due to either roof pendants of surrounding magnetite-rich migmatized sediments or mafic plugs having a greater magnetic content than the granite. Each of these anomalies is situated over a circular bog-filled depression identified from aerial photographs, and both were investigated on the ground in 1985 (D. Evans, personal communication, 1989). Although granite forms the only outcrop around the perimeter of these bog-filled depressions, their associated magnetic expression is thought to indicate differential weathering of the more strongly magnetic material.

The linear nature and the sharp magnetic contrast of the Overflow Pond Granite with the country rock along its northern boundary, suggest a steep or near vertical contact and that the thickness of the granite here, is considerable. In contrast, the jagged appearance of the southern boundary suggests that at this contact the pluton dips northward at a shallow angle. The northeastward continuation of linear magnetic units located to the south into the granite (dashed lineations immediately north and northeast of 'J' of Figure 3a) indicates that the granite is relatively thin along this southern boundary.


57°15'
56°30'
48°45
48°45'
57°30'W
48°30'N
48°30'
56°30'
48°15'N
48°15'N
57°30'W
56°45'W
AREA 1
AREA 2
-N-
A
A
A
B
B
C
D
E
F
G
H
H
I
I
J
K
L
M

Figures 2a and b. *Enhanced image results and geological sketch map, respectively for the southern part of the Victoria Lake Group, central Newfoundland (areas 1 and 2, Figure 1).* **Figure 2a:** *Grey-tone shaded relief image mosaic of high-resolution aeromagnetic total-field data. Note the lineament between the arrows. Features labelled (also discussed in text): A–Victoria Lake Group volcanics and sediments; B–gabbro; C–Snowshoe Pond granite; D–Buck Lake granite; E–Spruce Brook Formation metasediments, migmatites and granite of the Meelpaeg Subzone; F–Valentine Lake quartz monzonite; G–Rodeross Lake gabbro; H–Rogerson Lake Conglomerate; I–Topsails Intrusive Suite; J–Skidder basalt; K–Carboniferous subbasin; L–Harbour Round Formation; M–Victoria Lake Group?*

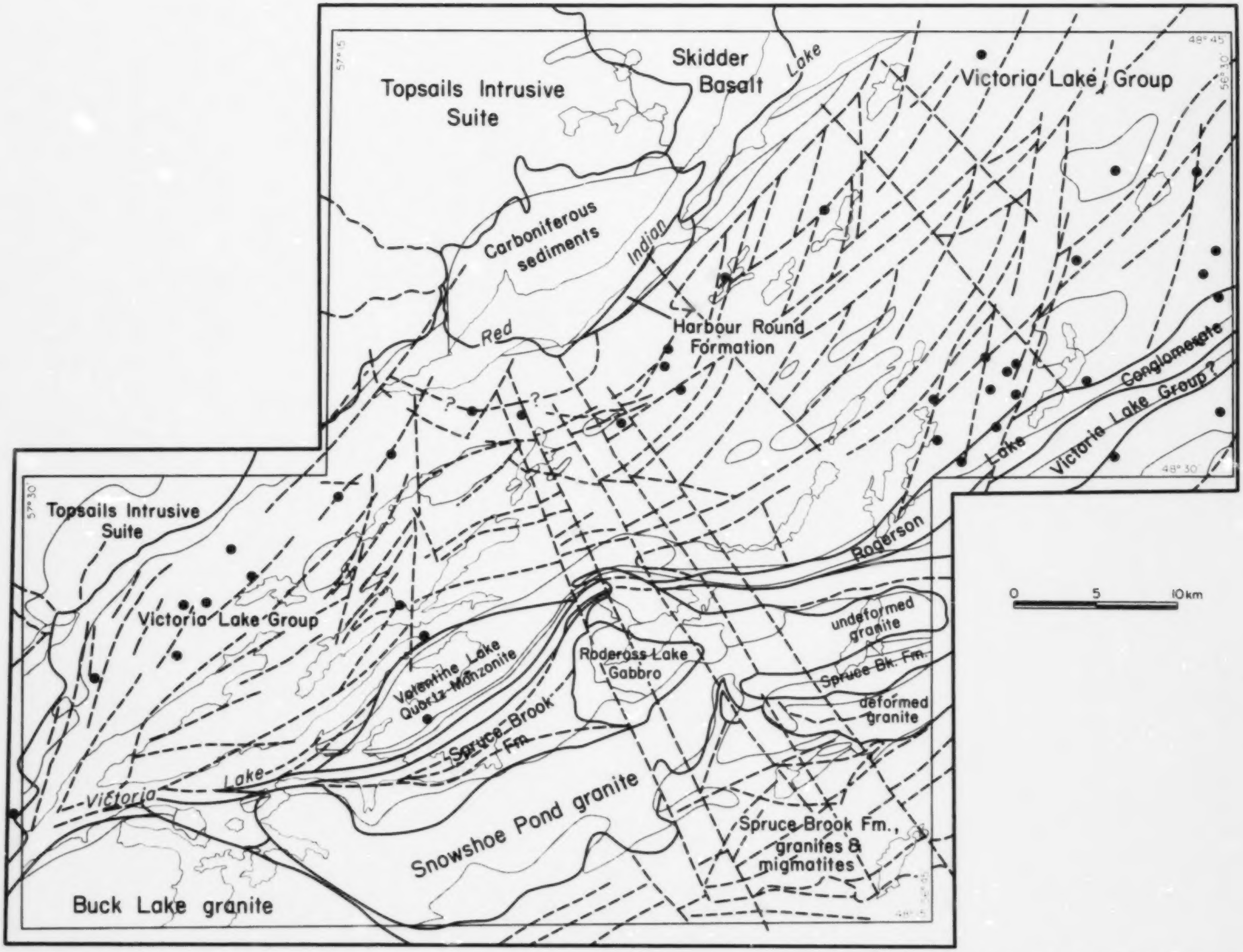
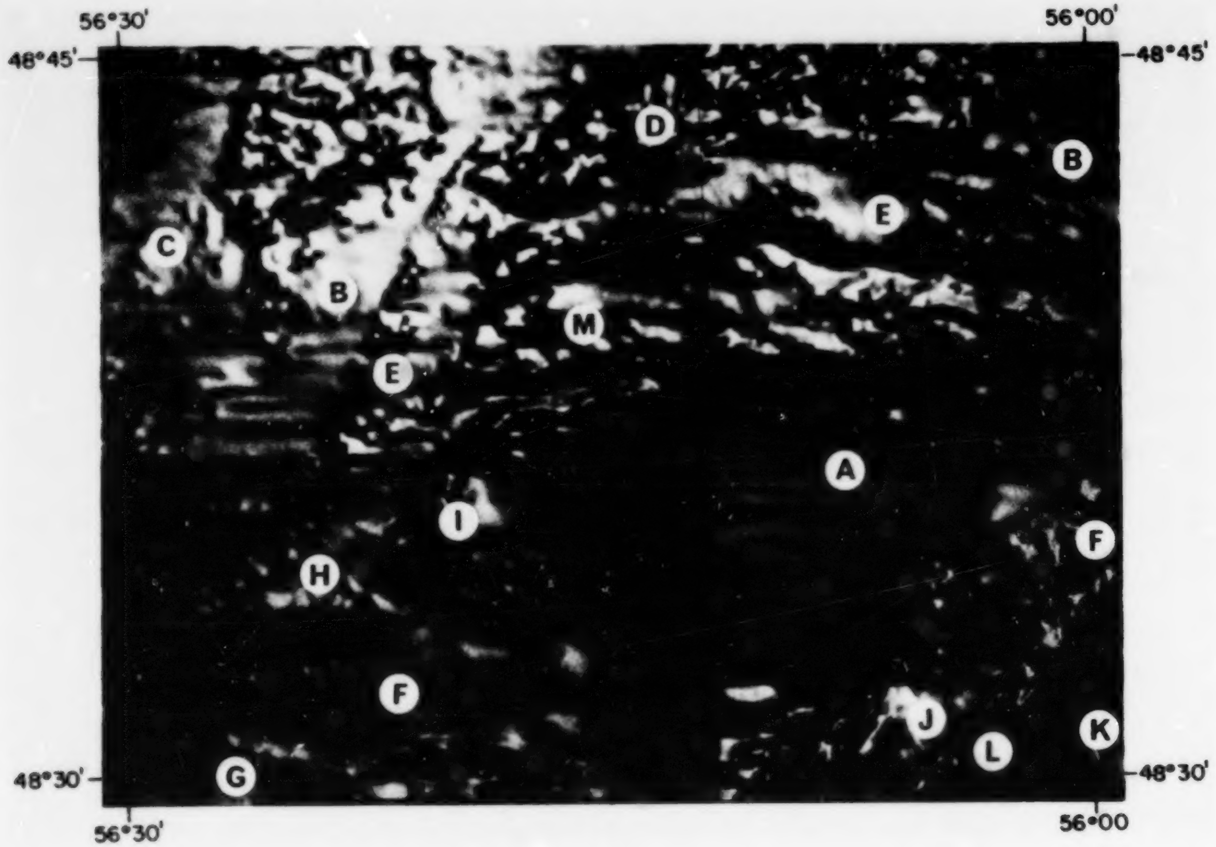

Topsails Intrusive Suite
Skidder Basalt
Lake
Victoria Lake Group
Carboniferous sediments
Indian
Red
Harbour Round Formation
Conglomerate
Victoria Lake Group?
Lake
Rogerson
?
?
Topsails Intrusive Suite
Victoria Lake Group
Valentine Lake Quartz Monzonite
Rodeross Lake Gabbro
undeformed granite
Spruce Bk. Fm.
deformed granite
Spruce Brook Fm.
Lake
Victoria
Snowshoe Pond granite
Spruce Brook Fm., granites & migmatites
Buck Lake granite
0
5
10km
48°45'
56°30'
48°30'
57°15'
57°30'
48°15'

Figure 2b: *Solid and dashed lines are magnetic unit boundaries and lineations, respectively. Lighter lines are outlines of selected geological units (from Kean, 1977a, 1977b, 1978, 1979, 1982; Kean and Jayasinghe, 1980; Colman-Sadd, 1987, 1988). Larger dots mark locations of known gold and massive sulphide mineralization (Kean and Evans, 1988; D. Evans, personal communication, 1989).*



Figures 3a and b: *Enhanced image results and geological sketch map, respectively, for Noel Paul's Brook (12A/9) map area, (area 3 of Figure 1); this area adjoins the upper east boundary of the area covered by Figure 2.* **Figure 3a:** *Grey-tone shaded relief image mosaic of high-resolution aeromagnetic total-field data. Features labelled (also discussed in text): A—Overflow Pond Granite; B—Victoria Lake Group volcanics and sediments; C—gabbro; D—Crippleback Lake Quartz Monzonite; E—Rogerson Lake Conglomerate; F—metasediments of the Meelpaeg Subzone; G—granite; H—Carter Lake Formation volcanics; I—gabbro; J—Great Burnt Lake granite; K—Pipestone Pond complex ultramafic rocks; L—Baie d'Espoir Group volcanogenic sediments; M—Pine Falls Formation volcanics.*

The northwestern portion of Figure 3a (near 'B') shows the trends of the more magnetic rich mafic volcanics within the volcanic and sedimentary package of the Victoria Lake Group. As in Figure 2a, the prominent northeast trends of units in the Victoria Lake Group are locally interrupted by faults trending north-northeast to north, as indicated by dashed lines in Figure 3b. These faults, associated with later deformation stages and flexuring, are most apparent near the northwest corner of the figure, where members of the Victoria Lake Group have the largest magnetic contrast. Two small gabbro bodies underlying areas in the northwest corner of Figure 3b (indicated by dotted lines north of position 'C' on Figure 3a) do not have a discernible magnetic expression. However, based on the magnetic anomaly at position 'C', a large gabbro body is interpreted to underlie this area.

The Crippleback Lake Quartz Monzonite, located at the northern edge of Figure 3a (at 'D'), exhibits a variable magnetic character indicating a strong magnetic zonation, probably reflecting a foliation. The boundaries of the monzonite defined by geophysical expression are mainly coincident with those mapped (Figure 3b). Lenticular mafic and felsic volcanic-rich units of the Victoria Lake Group that border the pluton tend to complicate its magnetic signature.

In this map area, the Rogerson Lake Conglomerate, identified by 'E' in Figure 3a, separates members of the Victoria Lake Group ('B') from volcanic rocks of the Pine Falls Formation ('M') to the southeast (Kean and Jayasinghe, 1980). The geophysically interpreted boundaries of the conglomerate are locally not coincident with those mapped on the ground (Figure 3b), and suggest that the unit is more consistent in width than surficial mapping would indicate.

The southwestern corner of Figure 3a is underlain by metamorphosed sediments (indicated by 'F') included in the

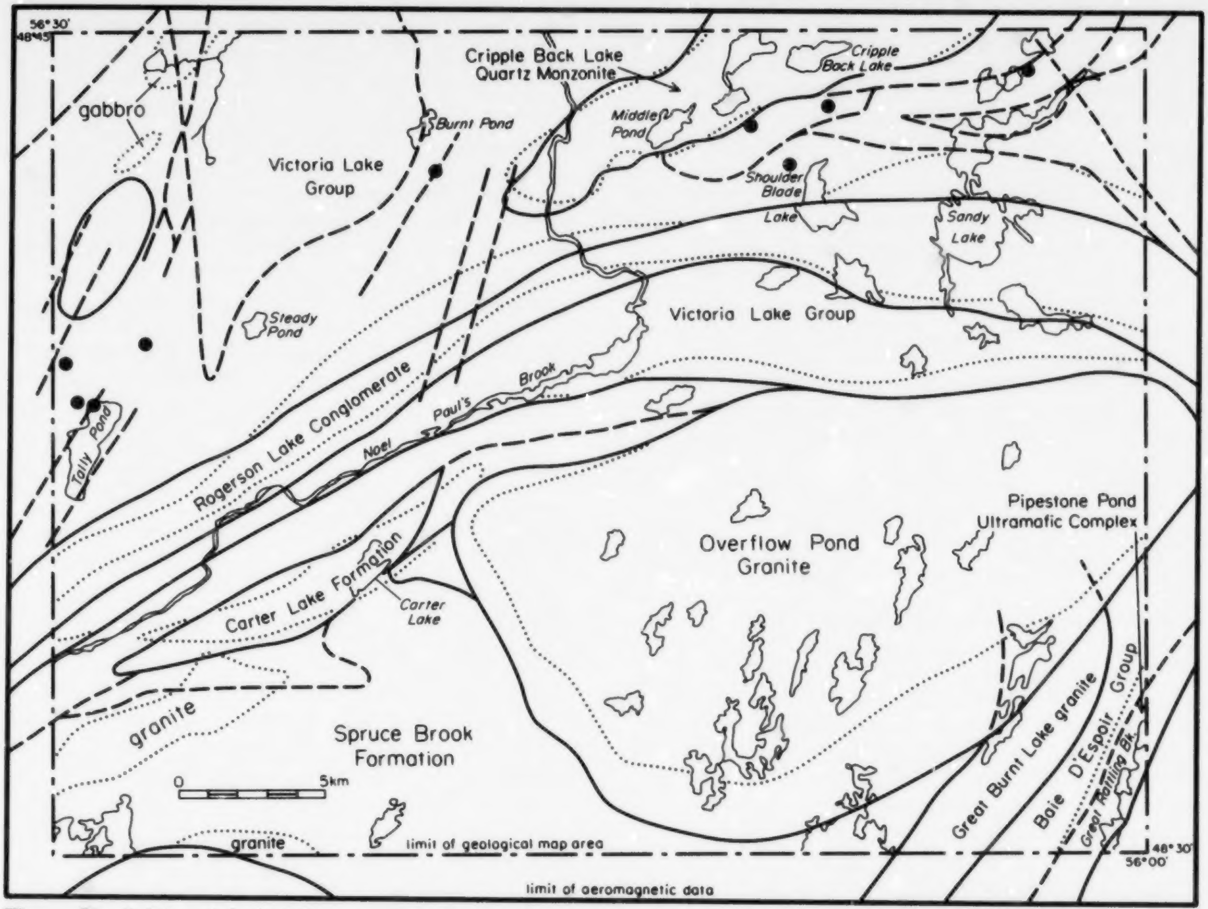

56°30'
48°45'
gabbro
Cripple Back Lake
Quartz Monzonite
Cripple
Back Lake
Burnt Pond
Victoria Lake
Group
Middle
Pond
Shoulder
Blade
Lake
Sandy
Lake
Steady
Pond
Victoria Lake Group
Rogerson Lake Conglomerate
Paul's Brook
Noel
Tally Pond
Pipestone Pond
Ultramafic Complex
Overflow Pond
Granite
Carter Lake Formation
Carter
Lake
granite
Spruce Brook
Formation
0
5km
granite
limit of geological map area
limit of aeromagnetic data
Great Burnt Lake granite
Baie D'Espoir Group
Great Rattling Bk.
48°30'
56°00'

Figure 3b: *Solid and dashed lines are boundaries of magnetic units and lineations, respectively. Dotted lines are outlines of geological units selected from Kean and Jayasinghe (1980); Jayasinghe (1979). Larger dots mark locations of known gold and massive sulphide mineralization (D. Evans, personal communication, 1989).*

Meelpaeg Subzone of Williams *et al.* (1988), who correlate them with probable Spruce Brook Formation equivalents mapped to the south and southwest in the Great Burnt Lake and Snowshoe Pond map areas, respectively (Colman-Sadd, 1985, 1988). The magnetic character varies from a relatively smooth field in the east to a variable field in the west, probably reflecting the increasing presence of mafic dykes to the west, as noted by Colman-Sadd and Swinden (1984a). Although two granite bodies have been mapped within the metasediments (Jayasinghe, 1979), only the southern pluton ('G' in Figure 3a) shows a discernible geophysical signature.

A lens of volcanic rocks ('H' in Figure 3a), the Carter Lake Formation (Kean and Jayasinghe, 1980), occurs adjacent to the metasediments. The geophysically defined boundaries of the unit differ slightly from the mapped boundaries (Figure 3b). A small gabbroic body may be present on the southeast flank of the Carter Lake Formation ('I' in Figure 3a) on the basis of correlation of its magnetic signature with those of gabbroic bodies elsewhere (cf. Snowshoe Pond granite, Figure 2a). A narrow lens of gabbro and pyroxinite also borders the Overflow Pond Granite along its eastern extreme in the Miguels Lake (NTS 2D/12) map area (Colman-Sadd and Russell, 1981, 1982).

The linear feature having a smooth magnetic field, located in the southeastern corner of Figure 3a at 'J', correlates with recently studied foliated, megacrystic, biotite granite outcrop (Colman-Sadd, personal communication, 1989) not identified in earlier studies (e.g., see Kean and Jayasinghe, 1980), which corresponds to a northward extension of the Great Burnt Lake granite, a linear granite body that transects the entire length of the Great Burnt Lake (12A/8) map area to the south (Colman-Sadd and Swinden, 1984a; Colman-Sadd, 1985). The strong northeast foliation to local mylonitization in this granite is not enhanced by this illumination angle. The linear magnetic anomalies that trace out its contacts with the Meelpaeg Subzone metasediments (Spruce Brook Formation?) to the northwest and Baie d'Espoir Group conglomerates and mafic volcanics to the southeast, may be followed into the Overflow Pond Granite by anomalies of decreasing magnitude. This granite is considered to be pretectonic, since, it is locally as strongly foliated as surrounding sedimentary units. In Figure 3b, the northward surrounding sedimentary units. In Figure 3b, the northward continuation of these anomalies into the granite is depicted by curvilinear dashed lines, which hint that the package of Great Burnt Lake granite and metasediments are warped by a fold, the axis of which is situated beneath the younger undeformed Overflow Pond Granite.

The extreme southeast corner of the map area ('K' in Figure 3a) is underlain by ultramafic rocks forming a northward extension of the Pipestone Pond complex units defined to the south in the Great Burnt Lake map area (Swinden, 1988). The serpentinites, pyroxenites and gabbros that comprise the complex define the western boundary of a tectonic klippen, in which rocks of the Gander Zone are exposed through those of the Dunnage Zone (Colman-Sadd and Swinden, 1984b). On the shaded relief image of Figure 3a, these ultramafic rocks contrast with the mafic volcanics and conglomerates of the Baie d'Espoir Group (Colman-Sadd and Swinden, 1984a) and Meelpaeg Subzone metasediments (Colman-Sadd and Russell, 1981, 1982), which are situated to the immediate northwest ('L' in Figure 3a), by their uniform magnetic signature and by an elevated level of magnetic field strength (Geological Survey of Canada, 1985c). The northwest boundary of the ultramafic units is marked by a large negative anomaly in Figure 3a, much as expected to the northwest of rocks of high magnetic susceptibility in the presence of the steeply inclined field of higher magnetic latitudes. A subtler northeast-oriented magnetic anomaly, situated just north of the Pipestone Pond complex, separates Baie d'Espoir Group volcanic-rich units to the southeast from the metasediments on its northwest side.

Known massive sulphide and/or gold occurrences are identified by dots in Figure 3b (D. Evans, personal communication, 1989). The good correlation of mineralization associated with north-northeast oriented linears within the Victoria Lake Group (Kean and Evans, 1988), particularly at Tally Pond and near Burnt Pond, is noteworthy (Figure 3a,b). These linears are interpreted as northwest-directed thrust faults, which displace interleaved mafic and felsic volcanic-rich members of the Victoria Lake Group. Mineralization also appears to be associated with magnetic linears located near the northern boundary of Figure 3b, on the southeast margins of the Crippleback Lake Quartz Monzonite.

FUTURE STUDIES

A methodology for combining the various geophysical datasets directly has, as yet, not been developed. As well, the VLF-EM datasets, quadrature and total-field, have not been incorporated into the study. Since each data type expresses a different characteristic of surface rocks, the potential existing in the proper combinations of these regional data types has not been exploited. The geophysical datasets can be digitally integrated with geochemical, digital topographic and digital geological datasets and models to further delineate areas of good economic potential. Such studies will be extended to the north of the study area by application of image enhancement techniques to high-resolution aeromagnetic, gradiometer and VLF-EM data, recently released for the Notre Dame Bay region, area 5 of Figure 1.

Recently, the manipulation and display of VLF-EM data from the GSC Skyvan survey of south-central Newfoundland has commenced. Colour profile plots of VLF quadrature, total-field and Fraser-filtered quadrature values have been produced. Aside from cultural features and the conductive coastline anomalies due to salt water, several conducting lithologies have been identified. Of particular interest are a large conductor near Gander Lake, coherent conductors within the Ackley Granite, and those within St. Joseph's Cove Formation sediments (Dickson, 1986) of the Baie d'Espoir Group in an area of extensive overburden.

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MINERAL OCCURRENCE DATA SYSTEM

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Mineral Deposits Section

The Mineral Occurrence Data System (MODS) is a two-part project, which consists of a manual Mineral Inventory File and a computerized Mineral Index. The project is designed to offer an efficient information service on all the mineral occurrences in Newfoundland and Labrador.

The Mineral Inventory File is part of the National Mineral Inventory and consists of mineral-occurrence descriptions that summarize all data on known mineral occurrences in the province. The file presently contains approximately 3,500 descriptions, including all mineral occurrences for the Island of Newfoundland (Figure 1) and selected areas in Labrador (Figure 2). The file and topographic location maps have been microfiched and duplicates are available upon request.

At present, the Mineral Inventory File for the Island of Newfoundland is being updated. This file was started in 1978, and since then, new geological mapping and exploration have been carried out and many new occurrences have been discovered. Updates were done in the past only on land tenure and major mineral occurrences. Areas that have been updated with new mineral occurrence descriptions include major parts of the 12B and 12H map areas. In addition, industrial minerals for all of Newfoundland have been documented.

Mineral occurrence maps that have updated geological bases have been published at a 1:250,000 scale and are available upon request. In addition, selected areas have been published at 1:50,000 and 1:100,000 scales. These maps contain locations, a listing and a brief description of the occurrences.

The computerized Mineral Index contains information selectively extracted from the Mineral Inventory File, which is stored on a Hewlett Packard 9000/560 Unix-operated mini computer, located at the Geological Survey of Newfoundland, Department of Mines, 95 Bonaventure Avenue, St. John's. The Geological Retrieval and Synopsis Program (GRASP) developed by the United States Geological Survey is used to manage and manipulate the database. GRASP permits users to perform complex searches of the mineral inventory data. This database presently contains information on approximately 3000 mineral occurrences. We plan to increase this to 3500 so that the computer index parallels the manual file. Plans are being made to publish this database on floppy disk along with a database program, which will enable users to install the database on their personal computers.

An index to the province's industrial mineral occurrences is also available. This database utilizes IBM compatible REFLEX software stored on a H.P. Vectra ES/12 micro computer and provides a quick index to industrial mineral occurrences. Users interested in a particular commodity can retrieve a listing of that commodity's occurrences, thus simplifying their search of the manual file.

The manual Mineral Inventory File is used as a quick reference to specific deposits. Exploration-company personnel frequently request information on a particular mineral showing or prospect. This is particular true of those companies new to the province or those moving into new areas. Generally, this preliminary research is done at the Department of Mines by government personnel. However, since a large part of the file has been microfiched, it is available to all companies at a reasonable cost.

The Mineral Inventory File is used daily by government geologists in land-use planning. Advice is given to various departments of government in establishing wilderness areas, hydro developments, provincial and national parks, and any other developments that may conflict with future mineral exploration and development. In addition, municipal councils and the Department of Municipal Affairs are advised of the location, extent, and nature of mineral deposits in specific areas, so that new housing and commercial developments, municipal parks, water reservoirs and sewage-disposal systems can be located, where possible, in areas of low-mineral potential.

During the past year, a questionnaire was circulated to mining and exploration personnel in an attempt to get some feed-back on the usefulness of the Mineral Occurrence Data System. We have received a good response and the results of the questionnaire, which are being tabulated, indicate that the system is of considerable help to the users. In the near future, a review of the system will be done, which will incorporate all useful suggestions made as a result of the questionnaire.

Copies of the file are made available to the various agencies of the federal government such as the Mineral Policy

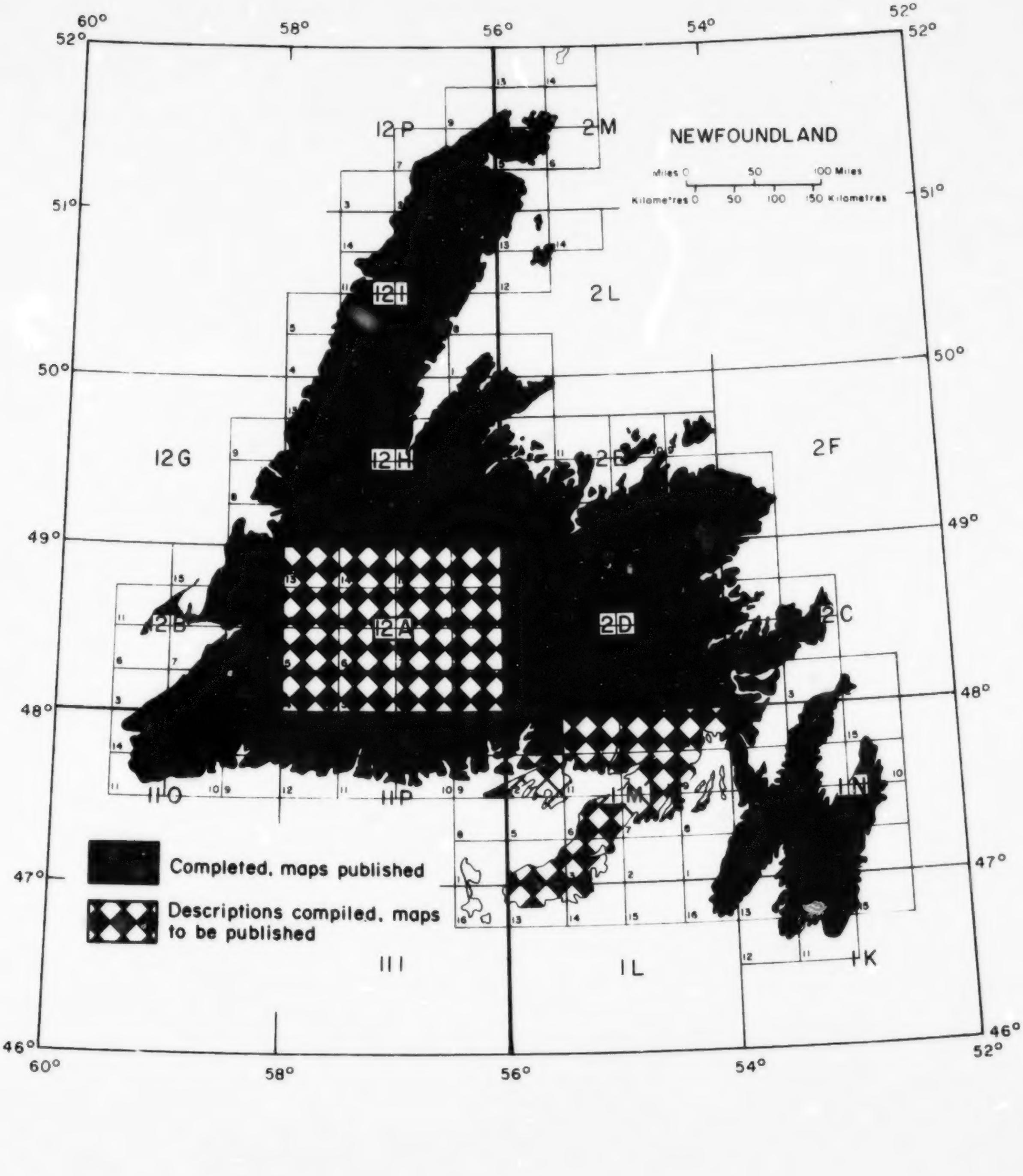
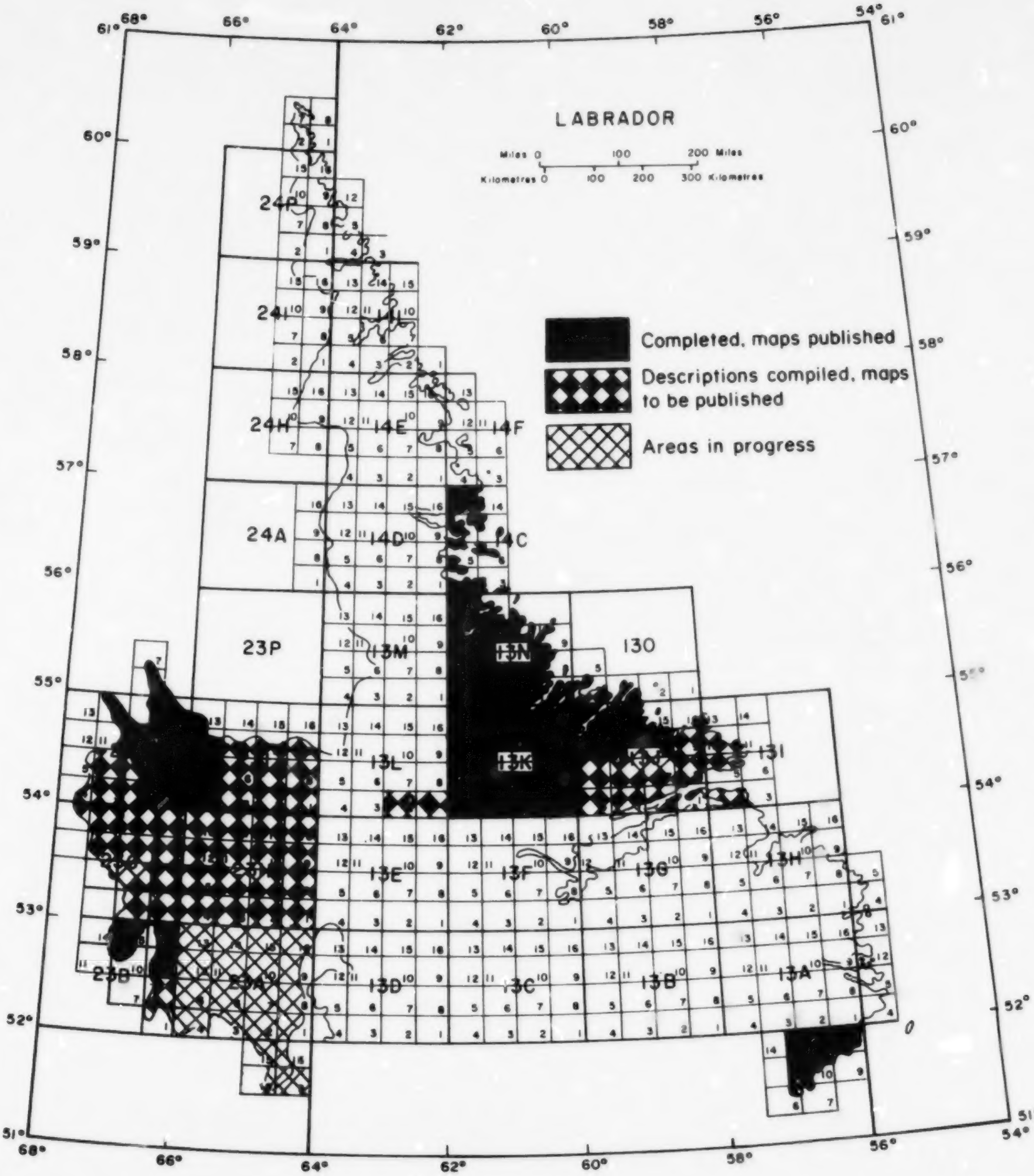

NEWFOUNDLAND
Miles 0 50 100 Miles
Kilometres 0 50 100 150 Kilometres
60°
58°
56°
54°
52°
52°
51°
50°
49°
48°
47°
46°
12P
2M
12I
2L
12H
12G
2E
2F
12A
2D
12B
2C
11O
11P
1M
1N
11I
1L
1K
Completed, maps published
Descriptions compiled, maps to be published

Figure 1. *Index map for Mineral Occurrence Data System project, insular Newfoundland.*


LABRADOR
Miles 0 100 200 Miles
Kilometres 0 100 200 300 Kilometres
Completed, maps published
Descriptions compiled, maps to be published
Areas in progress
24P
24I
24H
24A
23P
23B
23A
14E
14F
14D
14C
13M
13N
13O
13L
13K
13J
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68°
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58°
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57°
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53°
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51°

Figure 2. *Index map for Mineral Occurrence Data System project, Labrador.*

Sector and the Geological Survey of Canada. These are then adapted to the National Mineral Inventory, MINSYS and CANMINDEX.

The file is available to anyone who is researching mineral deposits in the province. It is useful to students who are writing papers on specific mineral deposits and to company and government personnel who are writing proposals for future work, or preparing information brochures about particular areas.

Using computers, the retrieval capability of the Mineral Index is optimized and complex retrievals can be made that would be virtually impossible if a manual system were used.

Current Research (1989) Newfoundland Department of Mines, Geological Survey of Newfoundland, Report 89-1, pages 297-299

CORE-STORAGE PROGRAM, 1988

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Mineral Lands Division
Mineral Resource Management Branch

INTRODUCTION

In 1979, the Newfoundland Department of Mines started a drill-core collection and storage program. The objective of this ongoing program is to provide well-maintained and representative collections of drill core from as many drilling programs as possible. This material would be made available, upon request, to the mineral exploration industry, government, university geoscientists and other geotechnical professionals, for further analyses and research.

This report summarizes activities during 1988, and presents an overview of the operations of the core-storage program.

DRILL-CORE ACQUISITION

During 1988, drill-core acquisition continued in insular Newfoundland; a total of 420,983 m of drill-core samples were added to the collection. Most of this core, 406,427 m, was acquired from the former Buchans co-tenancy of ASARCO−Abitibi-Price, and is from mine exploration drilling and other mineral exploration projects in the Red Indian Lake area. All this core material has been moved into the Buchans core library and carefully stacked on core-storage racks. The core material has been catalogued and a complete listing is presently being prepared for publication. In addition to the above, another 10,476 m of drill-core samples were delivered to our core libraries by exploration companies for confidential storage, and 4,080 m were collected by project personnel. Altogether, approximately 667,783 m of drill-core samples are now being stored in buildings operated by the Department.

A catalogue of all open core samples in Department of Mines core libraries is available from the Publications and Information Section of the Geological Survey Branch, Newfoundland Department of Mines.

DRILL-CORE LIBRARIES

The Department of Mines now operates six core libraries located at St. John's, Buchans, Springdale, Baie Verte, Pasadena and Goose Bay (Figure 1).

The St. John's library is staffed on a full-time basis except during the field season. It is equipped with a rock saw, a stereomicroscope, a magnetic susceptibility meter, a resistivity meter, an ultraviolet light and a McPhar TV-1 scintillometer.

The Goose Bay core library, which contains drill core from Labrador only, is not staffed full-time. It has facilities similar to, and shares equipment with, the St. John's core library.

The Pasadena core library is staffed full-time, has equipment similar to that of the St. John's library, and includes a ventilated room for storing radioactive core and a humidity-controlled room for storing salt core. The Pasadena building is filled to capacity with only a little working space made available for new core by moving some 'low priority' drill core outside the building for continued storage.

The Springdale library is operated by the resident Pasadena geologist from the Pasadena core library on an appointment basis only. It has a core splitter and a rock saw, but shares other geotechnical equipment with the Pasadena core library.

The Buchans core library contains drill-core samples from drilling undertaken by ASARCO and Abitibi-Price during the years these companies were active in mining and mineral exploration in Buchans and surrounding areas. This core library is not as yet fully operational, since it does not contain an office/core examination area and does not have staff assigned to it. Drillhole logs, location maps and other data on core in the Buchans library are being assembled in Pasadena, and from April 1989, all visits to the Buchans core library will be handled by staff from the Pasadena core library on an appointment basis.

The Department recently acquired a building in Baie Verte, suitable for use as a core-storage library. It is expected that this building will become operational during 1989 and will serve to store drill-core samples acquired from drilling projects on the Baie Verte Peninsula; this core library will also be managed from Pasadena.

FUTURE PLANS

During 1989, the core-storage libraries at Buchans and Baie Verte will become operational. Publication of a catalogue of drill-core samples in the Buchans core library will be completed early in 1989 and inspection of core samples in the building will be possible at the start of the 1989 field season.

Despite space problems in the Pasadena and Springdale core libraries, storage of confidential core in these buildings

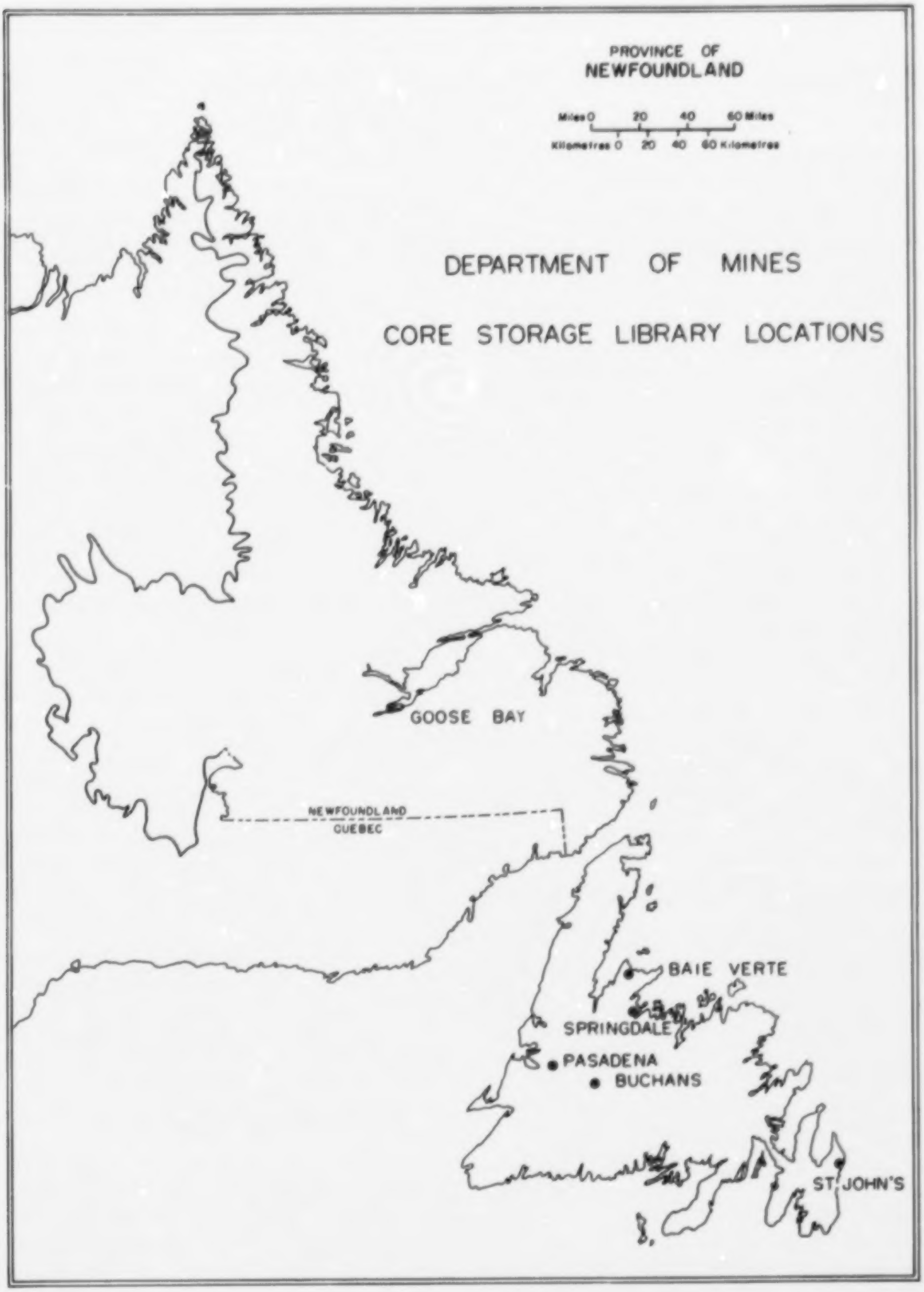
PROVINCE OF
NEWFOUNDLAND
Miles 0 20 40 60 Miles
Kilometres 0 20 40 60 Kilometres
DEPARTMENT OF MINES
CORE STORAGE LIBRARY LOCATIONS
GOOSE BAY
NEWFOUNDLAND
QUEBEC
BAIE VERTE
SPRINGDALE
PASADENA
BUCHANS
ST. JOHN'S

Figure 1. *Core libraries in the Province of Newfoundland and Labrador.*

is permitted. However, the rapid growth of core collections in all the Department core libraries and the continued high level of mineral exploration drilling, necessitates that the size of some core-sample collections must be reduced. The Department has doubled the number of core libraries it operates from three to six during the last two and a half years, whereas the size of its core collections has almost quadrupled. During 1989, reduction of some core-sample collections in the Pasadena core library will proceed.

Current Research (1989) Newfoundland Department of Mines,
Geological Survey of Newfoundland, Report 89-1, page 301

GEOSCAN–THE NATIONAL DATABASE FOR GEOLOGICAL INFORMATION: THE NEWFOUNDLAND CONTRIBUTION

Catherine Patey and Janet Gillespie
Publications and Information Section

GEOSCAN is a computerized bibliographic database of Canadian produced and/or Canada-related geoscience documents.

The Newfoundland Department of Mines is one of fifteen agencies that index documents for input into this database. The department's Mineral Assessment Report Library files and mineral inventory cards are the sources of the indexed records. Other agencies, e.g., the Geological Survey of Canada, also produce and index documents that concern this province.

In 1982, the National GEOSCAN Centre, which manages GEOSCAN, installed a new software system, called MINISIS, to input and retrieve the bibliographic data more effectively. In the implementation of the new system, all the old records were converted to a new format. For example, different types of keywords were moved into different fields: map types, illustrations, geological terms. Although most of the records did convert, they did not convert cleanly and completely. For example, records may be missing a title or authors and terms may have been assigned to the wrong fields. In cooperation with the National GEOSCAN Centre, these older records are being corrected.

All new records are input onto the new system via batch entry and are immediately retrievable, online. To date, only indexed records for assessment files are being keyed into the system, although other types of records, in particular, those for government publications, have been coded.

Besides indexing and keying new records, new coding sheets are being produced; the authority files or thesauri used in coding are being enlarged, and the terms used are being more clearly defined.

The non-confidential records in the 150,092 GEOSCAN records from across Canada, including those concerning this province, are available through online computer searches at the Geological Survey of Canada Library in Ottawa and in St. John's, at the Publications and Information Section. Searches are done on a cost-recovery basis with a minimal charge of $10.00 per search topic. A moderate size search (100 to 200 titles) costs between $15 and $45.

As well, GEOSCAN Newfoundland and Labrador indexes, (NTS, keywords, authors, both corporate and personal), current to February 1982, the date of the last update on the old system, are available from the Publications and Information Section. A computer printout index sorted by NTS values, to assessment files entered into GEOSCAN from 1982 to 1984, is also available.

This year two persons joined the GEOSCAN project–a geologist and a clerk. We have prepared a NTS index to assessment files received in the Publications and Information geofiles from April 1985 to June 1988. During the spring of 1989 other indexes from new and revised records in the database will be incorporated.

We welcome suggestions for improvements and appreciate being notified of any errors or discrepancies in the earlier indexes. This will aid us in making detailed modifications of the documents.

For further information on GEOSCAN, please contact Catherine Patey (709) 576-3159.

Preliminary lithological, petrological, and geochemical investigations of the Archean Florence Lake Group, central Labrador[1]

Terry D. Brace[2] and Derek H.C. Wilton[2]

Brace, T.D. and Wilton, D.H.C., Preliminary lithological, petrological, and geochemical investigations of the Archean Florence Lake Group, central Labrador; in Current Research, Part C, Geological Survey of Canada, Paper 89-1C, p. 333-344, 1989.

Abstract

The Florence Lake belt occurs in the most southerly portion of the Archean Nain Province, immediately north of the Proterozoic supracrustals of the Labrador Central Mineral Belt. The Florence Lake Group consists dominantly of mafic metavolcanic rocks with minor metasedimentary and metafelsic volcanic rocks. Ultramafic rocks, variably serpentinized and altered to talc-carbonate, occur as concordant intrusions within the group. The group was later intruded by the Archean tonalitic-trondhjemitic Kanairiktok Intrusive Suite. Geochemically, the mafic volcanics resemble tholeiitic ocean floor basalts, whereas the felsic volcanics are calc-alkaline. Proterozoic diabase and gabbroic dykes cut all other lithologies. Carbonatized ultramafics at the Baikie showing contain nickeliferous pyrrhotite with elevated PGE contents.

Résumé

La zone de Florence Lake se trouve dans l'extrême sud de la province archéenne de Nain, immédiatement au nord des roches supracrustales protérozoïques de la zone minérale Centrale du Labrador. Le groupe de Florence Lake se compose principalement de roches métavolcaniques mafiques, accompagnées de quelques roches métasédimentaires et roches volcaniques métafelsiques. Les roches ultramafiques, diversement serpentinisées et altérées en carbonate et talc, se présentent sous forme d'intrusions concordantes à l'intérieur de ce groupe. Ce dernier a plus tard été traversé par la suite intrusive archéenne de Kanairiktok, de caractère tonalitique et trondhjemitique. Géochimiquement, les roches volcaniques mafiques ressemblent aux basaltes tholéiitiques des fonds marins, tandis que les roches volcaniques felsiques sont calco-alcalines. Des dykes protérozoïques de diabase et de gabbro recoupent tous les autres types de roches. Les roches ultramafiques carbonatisées de la venue de Baikie contiennent de la pyrrhotine nickélifère caractérisée par une teneur élevée en éléments du groupe du platine (EGP).

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[1] Contribution to the Canada-Newfoundland Mineral Development Agreement 1984-1989. Project carried by the Geological Survey of Canada, Mineral Resources Division.

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The Grenvillian event: magmatism and high grade metamorphism

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Lithosphere and Canadian Shield Division

Emslie, R.F. and Hunt, P.A., The Grenvillian event: magmatism and high grade metamorphism; in Current Research, Part C, Geological Survey of Canada, Paper 89-1C, p. 11-17, 1989.

Abstract

One hallmark of the Grenvillian event is high grade regional metamorphism that reached peak intensity over a broad region about 1.1 Ga ago. The most widespread, characteristic plutonic igneous activity in the Grenville Province comprises rocks of anorthosite-mangerite-charnockite-granite (AMCG) suites. U-Pb zircon dates from sampling of the major complexes indicate they were intruded during at least three distinct periods at ~1.64, ~1.36, and ~1.16-1.13 Ga. The predominant younger magmatic episode, concentrated in the central part of the province, can be regarded as inaugurating the Grenvillian event which also included one or more subsequent intense thermal pulses, severe crustal deformation, uplift and cooling that affected much of the Grenville Province. High grade metamorphic terranes developed thermal peaks some 50-100 Ma after peak magmatic intensity, denoting a lag in regional heating of the crust that may be explained by existence of a thick, stabilized, subcontinental lithosphere prior to AMCG magmatism.

Résumé

L'un des faits marquants de l'épisode de Grenville a été le métamorphisme régional de forte intensité qui a atteint un maximum dans une région étendue il y a environ 1,1 Ga. L'activité plutonique ignée caractéristique la plus répandue dans la province de Grenville a donné les roches des suites anorthosite-mangérite-charnockite-granite. Les datations U-Pb faites sur le zircon lors d'un échantillonnage de reconnaissance des complexes intrusifs majeurs indiquent qu'ils ont été mis en place pendant au moins trois périodes distinctes il y a environ 1,64, 1,36 et de 1,16 à 1,13 Ga. L'épisode magmatique prédominant le plus ancien, concentré dans la partie centrale de la province, peut être considéré comme ayant amorcé l'épisode de Grenville qui a également englobé une ou plusieurs intenses pulsations thermiques ultérieures, une intense déformation de la croûte et un soulèvement suivi d'un refroidissement qui ont touché une grande partie de la province de Grenville. Environ 50 à 100 Ma après le maximum de l'intensité de l'épisode magmatique, des maximums thermiques se sont manifestés dans les terranes ayant subi un métamorphisme de forte intensité, ce qui indique un retard du réchauffement régional de la croûte qui peut-être expliqué par l'existence d'une épaisse lithosphère subcontinentale stabilisée, antérieure au magmatisme des roches des suites sus-mentionnées.

The Boundary Zone of the Nain-Churchill provinces in the North River-Nutak map areas, Labrador

I.F. Ermanovics, M. Van Kranendonk[1], L. Corriveau[2],
F. Mengel[2], D. Bridgwater[3], and R. Sherlock[4];
Lithosphere and Canadian Shield Division

Ermanovics, I.F., Van Kranendonk, M., Corriveau, L., Mengel, F., Bridgwater, D., and Sherlock, R., The Boundary Zone of the Nain-Churchill provinces in the North River-Nutak map areas, Labrador; in Current Research, Part C, Geological Survey of Canada, Paper 89-1C, p. 385-394, 1989.

Abstract

Proterozoic activity in Nain Province between 57°30'N and 58°00'N included emplacement of Napaktok dykes, crustal flexuring during deposition of Ramah and Mugford groups and development of breccia dykes along the conjugate Napaktok dykes fracture pattern. The easternmost distally located Proterozoic rocks in Nain Craton, show subgreenschist metamorphism and open fold styles that progress toward Churchill Province in Churchill Foreland Zone to amphibolite facies accompanied by two phases of folding, including isoclinal folds; rocks in the Boundary Zone attain granulite facies and are affected by transcurrent sinistral shear and by late, major, steep dip-slip faults.

Churchill Province (Tasiuyak gneiss terrane) also exhibits a progression of open to isoclinal folding and transcurrent shear, but all occurred at high grades including granulite facies. Structures in the Tasiuyak gneiss terrane are linked with structures in Proterozoic and Archean rocks of Nain Craton and formed in response to E-W transpression.

Résumé

Dans la province de Nain, l'activité au Protérozoïque entre 57°30' et 58°00' de latitude nord englobe la mise en place des dykes de Napaktok, un fléchissement crustal pendant l'accumulation des groupes de Ramah et de Mugford ainsi que la mise en place de dykes bréchiques le long du réseau de fractures conjuguées des dykes de Napaktok. Les roches protérozoïques les plus distales à l'est dans le craton de Nain présentent un métamorphisme du sous-faciès des schistes verts et des plis de style ouvert progressant, en direction de la province de Churchill dans la zone d'avant-pays de Churchill, vers un faciès des amphibolites accompagné de deux phases de plissement, incluant des plis isoclinaux; dans la zone de Boundary, les roches atteignent le faciès des granulites et ont été touchées par un cisaillement transversal à composante sénestre ainsi que par d'importantes failles d'effondrement tardives de fort pendage.

On remarque également dans la province de Churchill (terrain gneissique de Tasiuyak) une progression du plissement ouvert au plissement isoclinal et au cisaillement transversal, mais tous se sont produits pendant des épisodes de métamorphisme de degré élevé, incluant le faciès des granulites. Les structures dans le terrain gneissique de Tasiuyak sont reliées aux structures dans les roches du Protérozoïque et de l'Archéen du craton de Nain et se sont formées en réponse à une transpression E-O.

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Wakuach Gabbro sills of the Howse Lake area, western Labrador[1]

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Findlay, J.M., Fowler, T.D. and Birkett, T.C., Wakuach Gabbro sills of the Howse Lake area, western Labrador; in Current Research, Part C, Geological Survey of Canada, Paper 89-1C, p. 345-351, 1989.

Abstract

Gabbroic sills dominate the stratigraphy of the Howse Lake area in the central Labrador Trough. Sills were emplaced within a heterogeneous host rock sequence formerly correlated with the Menihek Formation, but here assigned to the Le Fer Formation. The sills commonly show a well developed internal stratigraphy defined by the size and frequency distributions of plagioclase phenocrysts and glomerocrysts. Although some sills are not glomeroporphyritic, mapping indicates that many contain both non-glomeroporphyritic and strongly glomeroporphyritic horizons, which are symmetrically developed with respect to sill contacts. Sharp, internal intrusive contacts indicate that some sills represent more than one magmatic injection. Fractional crystallization of plagioclase in a crustal feeder chamber and subsequent emplacement of crystal laden magma is thought to be the origin of glomeroporphyritic lithologies. Non-glomeroporphyritic sills are considered comagmatic, and may have been derived from the chamber prior to the onset of plagioclase fractionation.

Résumé

Les filons-couches de gabbro prédominent dans la stratigraphie de la région du lac Howse, au centre de la fosse du Labrador. Ces filons-couches ont été mis en place au sein d'une séquence hétérogène de roches encaissantes, que l'on avait autrefois corrélée avec la formation de Menihek, mais que l'on place ici dans la formation de Le Fer. Les filons-couches montrent généralement une stratigraphie interne bien développée, définie par la granulométrie et la distribution de fréquence des phénocristaux et des glomérocristaux de plagioclase. Certains filons-couches ne sont pas gloméroporphyriques, mais les travaux de cartographie indiquent que beaucoup d'entre eux contiennent à la fois des horizons non gloméroporphyriques et d'autres fortement gloméroporphyriques, qui se sont formés symétriquement par rapport aux contacts avec les filons-couches. Des contacts intruisifs internes bien définis, indiquent que certains filons-couches correspondent à plusieurs injections magmatiques. On estime que la cristalisation fractionnée du plagioclase dans un réservoir magmatique crustal et la mise en place ultérieure du magma chargé de cristaux, sont à l'origine des textures gloméroporphyriques. On considère que les filons-couches non gloméroporphyriques sont comagmatiques, et qu'ils tirent peut-être leur source du réservoir magmatique à une époque avant le début du fractionnement des plagioclases.

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[1] Contribution to Canada - Newfoundland Mineral Development Agreement 1984-1989. Project sponsored by Geological Survey of Canada, Mineral Resources Division.

[2] Ottawa-Carleton Geoscience Centre, University of Ottawa, Ottawa, Ontario, K1N 6N5.

[3] Commission géologique du Canada, Centre Géoscientifique de Québec, Sainte-Foy, Québec, G1V 4C7.

Geology and petrology of the Handcamp gold prospect, Robert's Arm Group, Newfoundland[1]

Karen A. Hudson[2] and H. Scott Swinden[3]

Hudson, K.A. and Swinden, H.S., Geology and petrology of the Handcamp gold prospect, Robert's Arm Group, Newfoundland; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 93-105, 1989.

Abstract

The small polymetallic Handcamp prospect is hosted by Lower Ordovician Robert's Arm Group volcanic and sedimentary rocks in north-central Newfoundland. The gold-sulphide mineralization is restricted to a 50 m wide structurally complex zone dominated by volcanic and volcaniclastic rocks, red and ferruginous chert and argillaceous rocks, which have been variably altered to a quartz-sericite-pyrite-magnetite assemblage. Alteration and sulphide mineralization appear to be both synkinematic and postkinematic. The association of late Ca-Fe-Mg carbonates and silicates with sulphides and gold suggests that CO_2-rich fluids were instrumental in the mineralizing process. Preliminary geochemical data indicate that gold enrichment correlates with enrichment of Sb, As and Mo.

The above evidence suggests that the Handcamp prospect may have been generated through much later shear-related hydrothermal deposition, rather than by exhalative volcanogenic processes, as proposed by earlier workers. If true, this has important implications for exploration for additional gold mineralization in the Robert's Arm Group.

Résumé

Les roches volcaniques et sédimentaires du groupe de Robert's Arm, datant de l'Ordovicien inférieur, dans la partie nord-centrale de Terre-Neuve, renferme la petite zone de production polymétallique possible de Handcamp. La minéralisation en or et sulfures se limite à une zone structurale complexe de 50 m de large, principalement occupée par des roches volcaniques et volcanoclastiques, des cherts rouges et ferrugineux et des roches argileuses qui ont été diversement altérées en une association de quartz, séricite, pyrite et magnétite. Il semble que l'altération et la minéralisation sulfurée soient à la fois syntectoniques et post-tectoniques. L'association des carbonates et silicates tardives de Ca, Fe Mg avec les sulfures et l'or semble indiquer que des fluides riches en CO_2 ont contribué au processus de minéralisation. Des données géochimiques préliminaires indiquent que l'enrichissement en or correspond à l'enrichissement en Sb, As et Mo.

Les indices ci-dessus semblent indiquer que la zone de production possible de Handcamp a peut-être été créée par une sédimentation hydrothermale beaucoup plus tardive liée à une phase de cisaillement, et non par des processus exhalatifs d'origine volcanique, hypothèse auparavant proposée par d'autres chercheurs. Si tel est le cas, ceci aura d'importantes conséquences pour l'exploration d'autres minéralisations en or du groupe de Robert's Arm.

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[1] Contribution to the Canada-Newfoundland Mineral Development Agreement 1984-1989. Project carried by the Geological Survey of Canada, Mineral Resources Division.

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Aeromagnetic total field, gradiometer, and VLF-EM survey of part of the Dunnage Zone, central Newfoundland[1]

J. Tod and E.E. Ready
Geophysics Division

Tod, J. and Ready, E.E., Aeromagnetic total field, gradiometer, and VLF-EM survey of part of the Dunnage Zone, central Newfoundland; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 23-27, 1989.

Abstract

An aeromagnetic total field, gradiometer, and VLF-EM survey was carried out over part of the Dunnage Zone in central Newfoundland. This portion of the Dunnage Zone is geologically complex and well suited to mapping with aeromagnetics. Aerodat Limited of Toronto carried out the contract in 1986-1987, using a rotary wing aircraft.

Survey results are available at two map scales: 1:25 000 aeromagnetic total field and gradiometer contour maps and 1:50 000 magnetic anomaly and gradiometer colour interval maps. The VLF-EM total field and quadrature profiles are printed on the back of the 1:50 000 maps.

The aeromagnetic results substantiate and expand on the known geology in this part of the Dunnage Zone. The gradiometer colour maps in particular, show clearly defined intrusive features and major transcurrent faults. A series of folded sills, as well as a thrust sheet, can be identified in the middle and northern regions of the survey area, respectively.

Résumé

On a effectué dans ue partie de la zone de Dunnage dans le centre de Terre-Neuve un levé combiné aéromagnétique du champ total, gradiométrique et électromagnétique à très basse fréquence (VLF-EM). Cette partie de la zone de Dunnage est géologiquement complexe et se prête bien à la cartographie par levé aéromagnétique. En 1986-1987, la société Aerodat Limited de Toronto a procédé, à l'aide d'un hélicoptère, à l'exécution du contrat.

Il est possible d'obtenir les résultats du levé à deux échelles cartographiques: les cartes gradiométriques et aéromagnétiques du champ total en courbes de niveaux, à l'échelle de 1/25 000, et des cartes gradiométriques et des anomalies magnétiques avec intervalles en couleurs, à l'échelle de 1/50 000. Les profils VLF-EM du champ total et les profils en quadrature sont imprimés au verso des cartes à l'échelle de 1/50 000.

Les données aéromagnétiques confirment et ajoutent aux connaissances géologiques de cette partie de la zone de Dunnage. Les cartes gradiométriques en couleurs font notamment ressortir des éléments intrusifs bien définis et les grandes failles de décrochement. On peut également reconnaître dans le centre et le nord de la région à l'étude une série de filons-couches plissés, ainsi qu'une nappe de charriage.

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[1] Contribution to the Canada-Newfoundland Mineral Development Agreement, 1984-1989. Project carried by the Geological Survey of Canada, Geophysics Division.

Preliminary report on a classification of Newfoundland granitic rocks and their relations to tectonostratigraphic zones and lower crustal blocks[1]

Harold Williams[2], W. Lawson Dickson[3], K.L. Currie[4], John P. Hayes[3], and John Tuach[5]

Williams, H., Dickson, W.L., Currie, K.L., Hayes, J.P., and Tuach, J., Preliminary report on a classification of Newfoundland granitic rocks and their relations to tectonostratigraphic zones and lower crustal blocks; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 47-53, 1989.

Abstract

Newfoundland granitic rocks are separated into nine categories, each named after a type example as follows: Long Range, Holyrood, Round Pond-Cross Hills, Twillingate, Burgeo, Middle Ridge, Mount Peyton, Topsails and Ackley. Distinctive categories of granitic rocks occur within discrete tectonostratigraphic zones or coincide with lower crustal blocks.

Precambrian plutons (Long Range, Holyrood) that predate the Appalachian orogenic cycle occur on opposing sides of the orogen in the Humber and Avalon zones, respectively. Late Precambrian to Early Cambrian plutons (Round Pond-Cross Hills) in the Humber and Avalon zones are interpreted as synrift with respect to the Appalachian cycle. Deformed early Paleozoic plutons of oceanic or island arc affinity (Twillingate) relate to constructive phases of the Dunnage Zone. Deformed middle Paleozoic plutons that follow metamorphic belts of the Gander Zone (Burgeo, Middle Ridge) probably relate to accretionary events. Other middle Paleozoic plutons (Topsails, Mount Peyton, Middle Ridge) that cross local zone boundaries occur on opposite sides of the Dunnage Zone and may reflect differences in lower crustal blocks. A few others (Ackley examples) cut boundaries of both zones and lower crustal blocks.

Résumé

Les roches granitiques de Terre-Neuve sont subdivisées en neuf catégories. Chacune est nommée d'après un exemple type: Long Range, Holyrood, Round Pond-Cross Hills, Twillinggate, Burgeo, Middle Ridge, Mount Peyton, Topsails et Ackley. Les catégories distinctives de roches granitiques apparaissent à l'intérieur de zones tectonostatigraphiques discrètes, ou coïncident avec des blocs de la croûte inférieure.

Les plutons précambriens (Long Range, Holyrood) mis en place avant le cycle d'orogénèse des Appalaches occupent les versants opposés de l'orogène dans les zones de Humber et d'Avalon respectivement. Les plutons d'âge précambrien supérieur à cambrien inférieur (Round Pond-Cross Hills) dans les zones d'Humber et d'Avalon ont été interprétés comme constituant un fossé d'effondrement syntectonique dans le cadre du cycle de l'orogénèse appalachienne. Les plutons déformés, d'âge paléozoïque inférieur, d'affinités océaniques ou typiques des zones d'arc insulaire (Twillingate) se rapportent aux phases constructives de la zone de Dunnage. La mise en place des plutons déformés d'âge paléozoïque moyen qui suivent les zones métamorphiques de la zone de Gander (Burgeo, Middle Ridge) concorde probablement avec des épisodes d'accrétion. D'autres plutons d'âge paléozoïque moyen (Topsails, Mount Peyton, Middle Ridge) qui traversent les limites locales de zones occupent des versants opposés de la zone de Dunnage, et reflètent peut-être des différences dans la nature des blocs de la croûte inférieure. Quelques autres (exemples d'Ackley) recoupent les limites des deux zones et les blocs crustaux inférieurs.

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Tectonic relationships along the proposed central Newfoundland Lithoprobe transect and regional correlations[1]

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Williams, H., Piasecki, M.A.J., and Colman-Sadd, S.P., Tectonic relationships along the proposed central Newfoundland Lithoprobe transect and regional correlations; in Current Research, Part B, Geological Survey of Canada, Paper 89-1B, p. 55-66, 1989.

Abstract

The proposed Lithoprobe transect of central Newfoundland, Meelpaeg Transect, crosses the Humber, Dunnage, Gander and part of the Avalon tectonostratigraphic zones. The transect also crosses the boundaries between Grenville, Central and Avalon lower crustal blocks. Clarification of surface zone boundaries, related kinematic studies and appraisals of metamorphic and plutonic zonal expressions provide insight in deciphering relationships between surface rocks and the geometry of lower crustal blocks.

Résumé

Le projet de transect Lithoprobe du centre de Terre-Neuve, le transect Meelpaeg, traverse les zones tectonostratigraphiques de Humber, de Dunnage et de Gander ainsi qu'une partie de celle d'Avalon. Il traverse aussi les limites séparant les blocs crustaux inférieurs de Grenville, du Centre et d'Avalon. La clarification des limites des zones en surface, des études de cinématique connexes et des évaluations des expressions zonales de nature métamorphique et plutonique permettent de déchiffrer les relations qui existent entre les roches superficielles et la géométrie des blocs crustaux inférieurs.

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